

ICE-DAMMED LAKES AND
MARGINAL GLACIAL DRAINAGE
on Axel Heiberg Island — Canadian Arctic Archipelago

INAUGURAL-DISSERTATION

ZUR

ERLANGUNG DER PHILOSOPHISCHEN DOKTORWÜRDE

VORGELEGT DER

PHILOSOPHISCHEN FAKULTÄT II

DER

UNIVERSITÄT ZÜRICH

von

HANS ULRICH MAAG

aus Winterthur und Winkel (ZH)

1971

Ausgeführt bei den Herren

Prof. Dr. F. Müller (McGill Universität, Montreal)

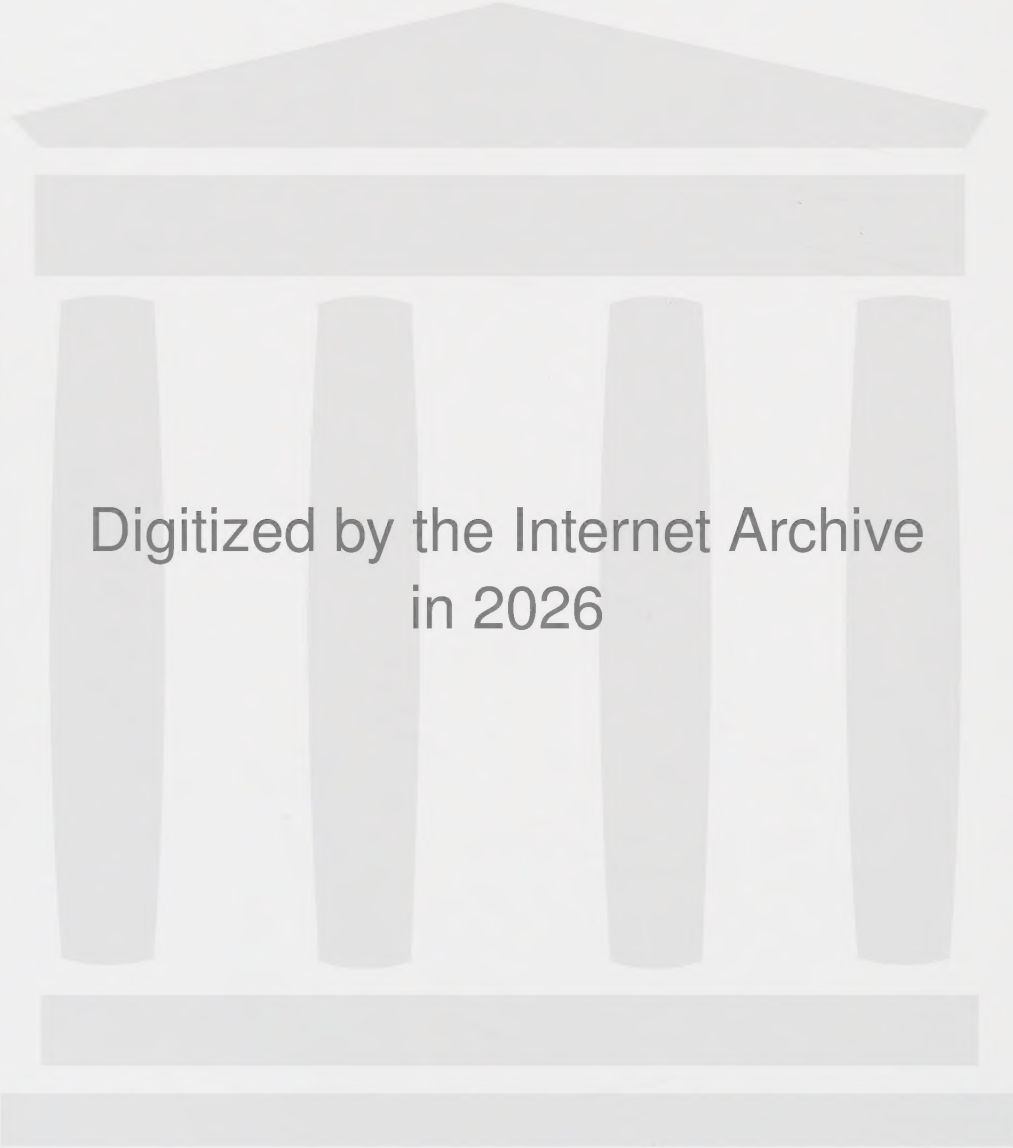
und Prof. Dr. H. Boesch (Universität Zürich),

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TO THE MEMORY OF MY PARENTS

ERRATA

- Page
- 21 Table 5, 1963, "16 Aug: still emptying" should read "16 July: 2nd collapse of outlet channel".
Table 5, 1st reference, "(Maag, p. 151)" should read "(Maag, 1963, p. 151)".
- 29 insert legend "Fig. 34: Phantom Lake sections and sites of sediment sampling and temperature measurements."
- 31 2nd paragraph, line 4, "which is the case every year" should read "which is not the case every year".
- 36 3rd paragraph, line 18, "Andrews, 1964" should read "Andrews, 1963".
- 42 Table 9, under Colour Lake, Surface, insert "1.0" against $\text{NO}_3(\text{mg/l})$, and against $\text{Cl}(\text{mg/l})$ " <0.01 " should read " <0.1 ".
- 51 5th paragraph, line 3, "emptying was lower" should read "emptying was slower".
- 67 2nd paragraph, line 3, "Mannerfelt, 1956" should read "Mannerfelt, 1945".
- 72 Fig. 91, line 2 of legend, "5 m" should read "5 km".
- 96 Table 12, under Sample No. 1, against $\text{Cl}(\text{mg/l})$, "0.04" should read "0.4".
- 99 5th paragraph, line 21, insert "to the downward percolation of water warmed by" between words "mainly" and "passing".
- 114 5th paragraph, line 3, insert "suggests that the amount of recent mass change" between words "it" and "varies".
- 123 4th paragraph, line 8, "quantitatively" should read "qualitatively".
- 136 2nd to last reference, "Heinsheimer, G.C." and "p. 33-38" should read "Heinsheimer, G.J." and "p. 24-32",
and add reference, "Heinsheimer, G.J., 1954b: Der Durchbruch des Morenogletschers, Lago Argentino, Patagonien, 1953. Zeitschrift für Gletscherkunde und Glazialgeologie, Bd. III, Heft 1, p. 33-38".
- 145 add reference, "Sissons, J.B., 1958b: Sub-glacial stream erosion in southern Northumberland. Scottish Geographical Magazine, Vol. 74, No. 3, p. 163-174".

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The photographs in this report were taken by the author except for the following ones:-

Nos 21, 26b, 42, 44, 69, 70, 86, 121, 122d, 127	by A. Maag,
Nos 7, 20a, 32b, 33, 45, 87, 118	by F. Müller
Nos 53, 54	by M. Müller-Vonmoos
Nos 41, 122c	by C.S.L. Ommanney
No. 135	by A.C.D. Terroux
Nos 8, 32a, 91, 139, 140, 141, 142	by RCAF

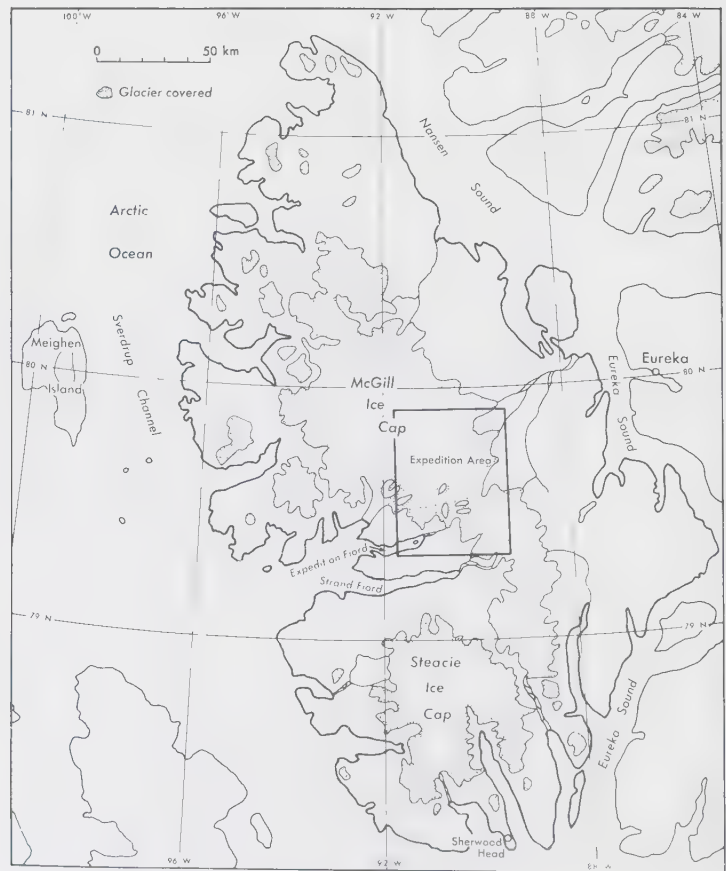
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(v)



Canadian Arctic Archipelago



Axel Heberg Island

Chapter A

INTRODUCTION

1. Scope of the Study

The main aim was to investigate the effect of glacial meltwater at the margins and in the tongue area of arctic valley glaciers with special reference to the rôle played by different types of ice-dammed lakes in the drainage régime of the rivers. This may further contribute to the understanding of geomorphological processes and forms in the immediate vicinity of present and former glaciers.

2. The Expedition

The fieldwork for this study was performed as part of the programme of McGill University's expedition to Axel Heiberg Island, 79°30' N, 90° W (Figs 1 and 2). The general aim of the expedition was "to study the evolution of the mountainous and strongly glacierized area of the central part of western Axel Heiberg Island" (Müller, 1961a, p.2). At the outset very little was known of the area so that, though the main emphasis was on glaciology, a number of other sciences, e.g. geology, geomorphology, geophysics, meteorology, photogrammetry, botany and zoology formed part of the integrated programme. This offered a unique opportunity of obtaining first hand information of neighbouring sciences, important in the interpretation of results. Starting in 1959 the work of the expedition continued through several consecutive summers. This continuity is very important in arctic research as the weather can differ very much from year to year.

At times, a light aircraft was available for transportation and reconnaissance flights. Excellent aerial photographs (Haumann, 1963, p. 86-89) and accurate large scale maps (Müller, 1963⁶) exist.

During the summer of 1960 (8 May - 31 August) the author was assistant to Professor Fritz Müller, the scientific leader of the expedition. The following year (12 May - 30 August 1961) and again in 1963 (19 June - 27 August) the opportunity was given him to pursue an individual research programme.

The working up of the material and the preparation of this report was done at the Department of Geography, University of Zürich, Switzerland, under the supervision of Professor H. Boesch.

3. Area of the Study

The Expedition Area, the approximately 900 km² selected for intensive study, lies in the mountainous central part of western Axel Heiberg Island (Fig. 2).

a) Glaciers

Several different types of ice bodies are present in the area: ice caps and ice fields, outlet glaciers, valley glaciers, cirque glaciers; some of the glaciers are advancing and some receding.

The glaciers belong to Ahlmann's "sub-polar" type (Ahlmann, 1935, p. 102, 103, 106). As the englacial temperatures of the investigated glaciers are negative, at least to a considerable depth (Müller, 1963h, p.87) and as the whole area belongs to the zone with continuous permafrost, the glacial meltwater drains in general on the surface and along the margins of the glaciers. There is only run-off in summer, i.e. from approximately the middle of June to the end of August.

Almost all the work reported here was done at the margins of the Thompson, White and Crusoe glaciers (Fig. 3), descriptions of which are given in Table 1.

In addition, some knowledge of the conditions at glacier and ice cap margins outside the Expedition Area was gained during several flights in a Piper Super-Cub on a short visit in 1960 to the most northern part of Axel Heiberg Island and from intensive study of aerial photographs.

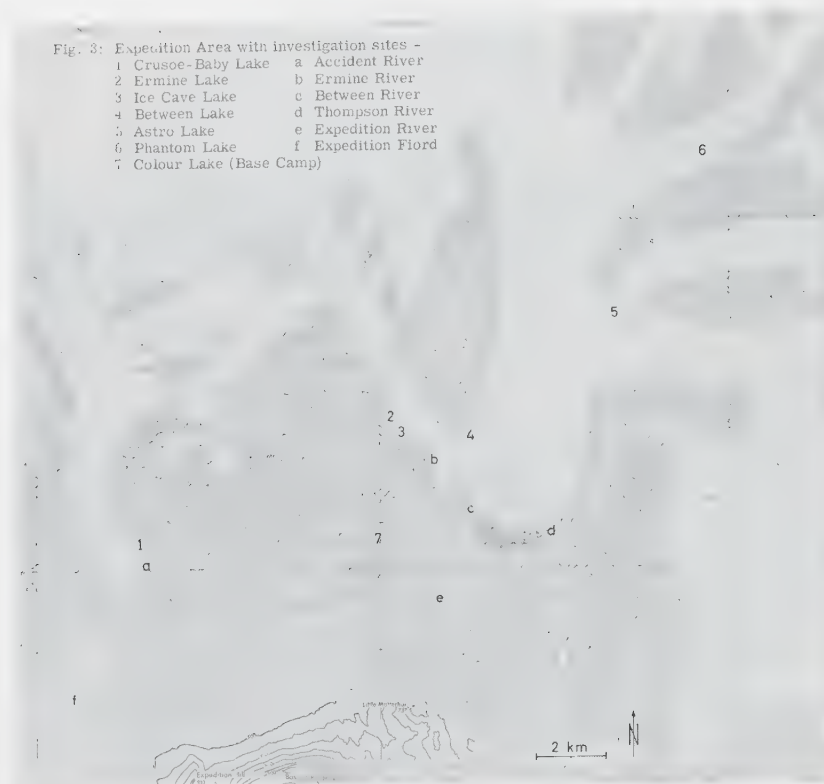


TABLE 1: The glaciers investigated.

	Thompson Glacier	White Glacier	Crusoe Glacier
Type	large outlet glacier of McGill Ice Cap	medium-sized, alpine-type valley glacier with separate accumulation basin	medium-sized, alpine-type valley glacier with separate accumulation basin
Orientation	N - S	NNW - SSE	N - S
Approx. size: area length width	230 km ² 38 km from ice-divide to snout 3 - 5 km	38 km ² 14.5 km 5 km in accumulation area 1 km in ablation area	43.5 km ² 13 km 1.5 km in tongue area
Average gradient: (ablation area)	3% (middle part 2%)	10% (with 3 steps)	9% (with 3 steps)
Snout: form elevation a.s.l. behaviour	cliff, some 40 m high 55-100 m advancing approx. 20 m/year	smooth 65 m receding approx. 5 m/year	cliff, 10-20 m high 110 m advancing approx. 10 m/year
Moraines	some distinct surface moraines; "push-complex" in front of snout	well developed lateral and terminal moraines	no well developed moraine ridges
Remarks	The Thompson and White glaciers are juxtaposed in the tongue area.		

Data partly obtained from Müller, 1961b, 1962a, 1963g and Ommanney, 1968.

b) Geology and Geomorphology

Triassic to early Tertiary sedimentary rocks found in the Expedition Area consist mainly of marine sandstones, siltstones and shales (Fricker, 1963a, p.5 ff). This sequence was deposited in the axial part of the Sverdrup Basin (Tozer, 1960, p.2,16).

Basalt flows as well as dykes and sills of gabbro and diabase, which have intruded the major part of the sedimentary rocks, occupy comparatively large areas because of their higher resistance to weathering (Fricker, 1962, Preliminary Geological Map; 1963b, p. 128).

In general the area has been strongly folded and faulted. Two systems of folding are recognized: the principal one probably of early Tertiary time trending north-northwest and a less pronounced one probably of a later period of deformation trending east-northeast (Fricker, 1963a, p. 132-144). The folds are generally characterized by broad synclines and narrow anticlines (Kranck, 1961, p. 148).

The intrusion of Upper Palaeozoic evaporites was contemporaneous with the Tertiary tectonic movements (Thorsteinsson and Tozer, 1960, p. 14-17). Diapirs are often found in the cores of anticlines, but evaporites have also intruded along fault planes (Fricker, 1963b, p.130; Hoen, 1964).

A northerly and an easterly striking fault system (mainly gravity faults) can be distinguished in the Expedition Area (Fricker, 1962, Preliminary Geological Map; 1963b, p.130). The main valleys, e.g. the Thompson Glacier and the White Glacier valleys, are to a high degree structurally controlled (Fricker, 1963a, p. 138-140).

The whole area is underlain by permafrost estimated to be 600 m thick. This is one of the main reasons why the mountainous landscape shows - despite its generally arid appearance - a remarkable amount of signs of fluvial activity (Rudberg, 1963b, p.139, 146-148). The meltwater and the water from sporadic heavy rainstorms cannot penetrate the ground but runs off almost immediately as surface water.

Striae, roches moutonnées, till and erratica (signs of former more extensive glacierization), solifluction, patterned ground features, mudflows, wind action forms, as well as other small geomorphological landforms were studied and mapped in detail by Rudberg (1963b, p.141-150). Evidence for at least two former glacierizations has been found in the Expedition Area (Boesch, 1963, p. 165). The age of fossil roots and drift wood and of marine shells found on old emerged beaches and the interpretation of a peat deposit situated some 2 km from the present snout of the Thompson Glacier prove that during the last 9000 years the glaciers have never advanced more than a few hundred metres beyond the present positions and that some 6000 years B.P. there was a "postglacial optimum" when a larger area was ice-free than today (Müller, 1966, p.155).

c) Climate and Vegetation

From Table 2 it can be seen that the weather conditions (and consequently the position of the equilibrium line and the duration of the ablation period) can vary widely from year to year but that in general the climate appears to be a little less cold and dry than that of the nearest weather stations, Eureka and Isachsen.

Despite the northerly latitude, with its short vegetation period and the relatively dry climate, some 140 species of vascular plants (including the largest plant of the island, the arctic willow which creeps over the ground), about 200 mosses and some 250 species of lichens have been recorded (Beschel, 1961a, p.179-182; 1963b, p.199). But closed vegetation is nowhere extensive. Protection by vegetation is therefore of limited importance (Rudberg, 1963a, p. 214).

d) Selection of Lakes and Rivers for Investigation

The logistic problems of such a remote area limited the number of sites which could be investigated and the time available for the investigations. The study was therefore concentrated on a few typical ice-dammed lakes and glacial rivers relatively close to the Base Camp. The main focus was on the Between Lake - Between River - Expedition River system, while a lesser study was made of Crusoe-Baby Lake and the Accident River, of Ice Cave Lake and the Ermine River, and of the Phantom Lake system.

Table 2: Climatic summary

(a) Some climatological data from the Base Camp weather station (79° 25'N, 90° 45'W, 190 m a.s.l.) for the four field seasons 1960, 1961, 1962 and 1963, in comparison with the "long-term" weather data from Eureka and Isachsen.

	MAY (19-31)		JUNE (23-30)				JULY				1 - 25 AUGUST				(1-31 Aug.)		
	1960	1961	1960	1961	1962	1963	1960	1961	1962	1963	1960	1961	1962	1963	1960	1961	
Temperature:	mean daily mean, °C	-5.9	-10.1	+4.3	-1.0	+4.8	+2.0	+6.6	+4.5	+8.6	+7.7	+5.7	+3.5	+7.9	+5.4	+4.5	+3.2
	highest max., °C	+1.7	-0.6	+17.2	+6.9	+12.2	+6.0	+15.8	+13.1	+18.9	+14.9			+18.9	+16.3	+16.7	+12.0
	lowest min., °C	-12.2	-18.1	-2.2	-12.2	-0.6	+0.6	-1.4	-1.4	+1.1	-0.8			+2.2	-5.7	-8.3	-2.1
	days with max. > 15°C	0	0	2	0	0	0	1	0	8	0			2	2	3	0
	days with min. < 0°C	13	13	10	24	1	0	1	6	0	2			3	3	12	7
Precipitation:	total, mm	16.0*	3.5*	2.5	10.9	0.25	1.5	61.8	98.5	18.6	7	12.2	19.0	13.8	1.5	17.0	?
	days, with 0.3 mm) or more	9*	2*	2	5	0	1	12	19	7	3			9	2	2	12
	days, with trace) more	11*	5*	8	13	5	5	16	23	11	7			12	6	6	15
Bright sunshine:	total, hour	101.1*	183.1*	362.5*	249.6*	74.0	67.5	236.9*	110.5*	357.3	444.5			248.6*	193.7*	243.7	37.0
	mean daily mean, hour	7.8*	15.3*	12.1*	8.3*	9.3	8.4	7.6*	3.6*	11.5	14.3			9.9*	7.8*	9.7	1.5
	% of possible	32.4*	63.6*	50*	35*	38	35	32*	15*	49	60			55*	42*	52	8
Cloud cover:																	
mean daily mean, tenths	8.5*	5.7	5.8	7.8	7.4	~7	7.1	9.0	7.3	5.0	7.0	7.6	5.6	7.3	6.0	8.9	7.0
																	8.1

(b) Equilibrium line, ablation area and mass budget of White Glacier

Budget year	Elevation of equilibrium line (m a.s.l.)	Max. extent of ablation area (% of total glacier surface)		Mass budget	
1959 - 1960	1200	53		negative (loss of ice mass)	
1960 - 1961	915	25		slightly positive	
1961 - 1962	1310	70		strongly negative	
1962 - 1963	~1200	~55		negative	
1963 - 1964	~400	~9		strongly positive	

* Data from Lower Ice Station, tongue of White Glacier.

** "Long-term" means from weather records of only 13 to 16 years' duration.

Eu. = Eureka, permanent weather station some 100 km northeast of Base Camp, 2 m a.s.l.

Is. = Isachsen, permanent weather station some 300 km south-west of Base Camp, 30 m a.s.l.

References: Andrews, 1964, Appendix;

Havens, 1961; Havens et al., 1965, Appendix;

Müller, 1963f, p. 61; 1966, p. 140 (Table 1), p. 146 (Table 2);

Müller and Roskin-Sharlin, 1967, Appendix.

Chapter B

INVESTIGATIONS OF ICE-DAMMED LAKES

1. The Ice-dammed Lakes in the Expedition Area

Location, size, type, ice cover, filling stage and drainage mechanism of all ice-dammed lakes in the Expedition Area have been recorded from observations on the ground, reconnaissance flights, and the study of maps and aerial photographs.

One difficulty was to distinguish between fossil lake basins (not included in this survey) and lake basins that were partially or totally emptied the summer prior to the date of the visit or of the photographs. Comparisons of photographs of different years (vertical air photographs of 1959 and 1960 and some oblique ones of 1961 and 1962) and of different days as well as a good number of field observations helped in most cases to solve this problem satisfactorily.

Air photographs covering the whole of Axel Heiberg Island have been studied in the Air Photo Library in Ottawa in order to make sure that the data collected in the relatively small Expedition Area, of which almost 60% is ice and firn covered, are representative.

a) Number and Size of Ice-dammed Lakes

The smallest lakes included in the survey have an area of some 500 m², the largest ones reach several million m². Some 125 ice-dammed lakes and lakelets were recognized in the Expedition Area, i.e. an average of some 24 lakes on every 100 km² of firn and ice. Most of the lakes are very small, approximately 10% smaller than 1000 m², 50% smaller than 5000 m², 65% smaller than 10000 m² and 90% smaller than 50000 m². Figure 4 points out the relative importance of the large lakes. For comparison the ice-dammed lakes in the region of the Aletsch Glacier and its tributaries (temperate glaciers in the Swiss Alps) have been recorded from observations on the ground, study of aerial photographs and the 1:10 000 topographical map (Eidg. Landestopographie, 1960, 1962, 1964). It was found that in the Aletsch Glacier region the ice-dammed lakes are less numerous and much smaller (Table 3); all except two being smaller than 1000 m², so that the average total lake area per unit investigation area is more than 100 times less than that of the Expedition Area.

In his survey of ice-dammed lakes in Alaska, Stone (1955, 1963b) only included those over approximately 400 000 m²; he cited 53 such lakes in an area with a glacierized part some 80 times larger than that of the Expedition Area (Table 3). He states that this is probably one of the highest concentrations in the world (Stone, 1963b, p. 332). In the Expedition Area there are 4 lakes larger than 400 000 m², which means that in the central part of Axel Heiberg Island (and this is true for the other mountainous and glacierized parts of Axel Heiberg Island also) the concentration of large ice-dammed lakes is at least as high, if not higher, than in Alaska with the possible exception of the very largest lakes.

Table 3: Ice-dammed lakes in the Expedition Area, Axel Heiberg Island, Alaska and the Swiss Alps

	Expedition Area	Axel Heiberg Island	Alaska (area investigated by Stone, 1955, 1963b)	Aletsch Glacier area	Swiss Alps
Latitude	79°30' N	78°01'20" N	56° - 62° N	46° 30' N	46° - 47° N
Geophysical type of glacier	sub-polar	sub-polar	temperate	temperate	temperate
Approx. area (km ²) covered by ice & firn = 7 times Expedition Area	500 1	12 500 ⁽¹⁾ 25 x	40 000 ⁽²⁾ 80 x	136 ⁽³⁾ 0.27 x	1 500 ⁽⁴⁾ 3 x
Approx. No. of ice-dammed lakes & lakelets (> 500 m ²) = average No. of lakes per 100 km ² area covered by ice and firn	120 24	- -	- -	8 ⁽⁵⁾ 6 ⁽⁵⁾	- -
Approx. total surface area of all lakes (m ²) = % of total area covered by ice and firn	11 000 000 27	- -	- -	30 000 ⁽⁶⁾ 0.2	- -
Approx. total volume of water ponded in all lakes (m ³) approx. average amount of water per km ² area covered by ice and firn	550 000 000 1 000 000	- -	- -	120 000 ⁽⁶⁾ 900	- -
Total No. of large ice-dammed lakes (and average No. per 1000 km ² area covered by ice and firn):					
> 400 000 m ²	4 (8)	~ 25 (2.0)	53 (1.3)	0	0
> 1 000 000 "	2 (4)	-	38 (1.0)	0	0
> 2 000 000 "	1 (2)	~ 10 (0.8)	20 (0.5)	0	0
> 5 000 000 "	1 (2)	~ 2 (0.2)	8 (0.2)	0	0

(1) Müller, 1961b, p. 43; (2) Estimated with maps from Field (1958); (3) Kasser, 1961, p. 217; (4) Kasser, 1959, p. 155;
(5) Number depends much on season and year, some of them ponded by morainic material;
(6) Includes Märljensee which contains most of this water.

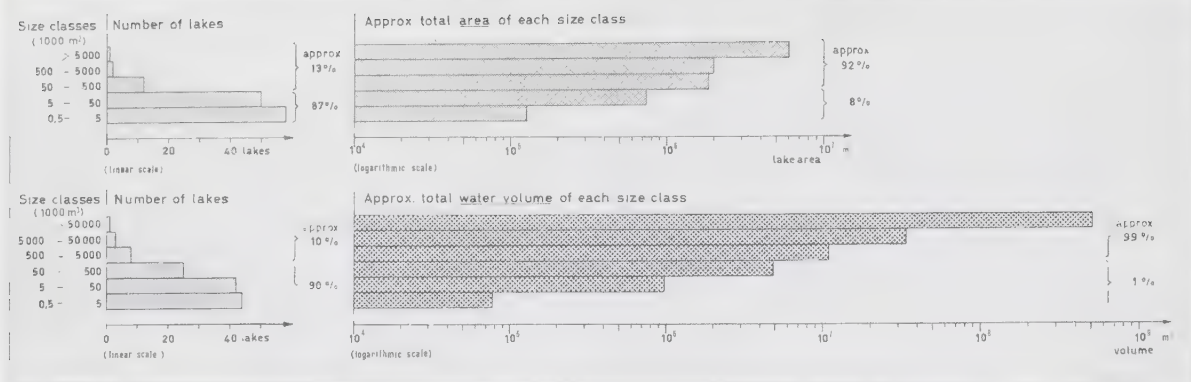


Fig. 4: Some statistics of the ice-dammed lakes.



Fig. 5: Types of ice-dammed lakes in Expedition Area (diagrammatic): A - supraglacial; B - marginal; C - at junction of converging glaciers; D - in tributary stream valleys; E - in tributary glacier valleys.

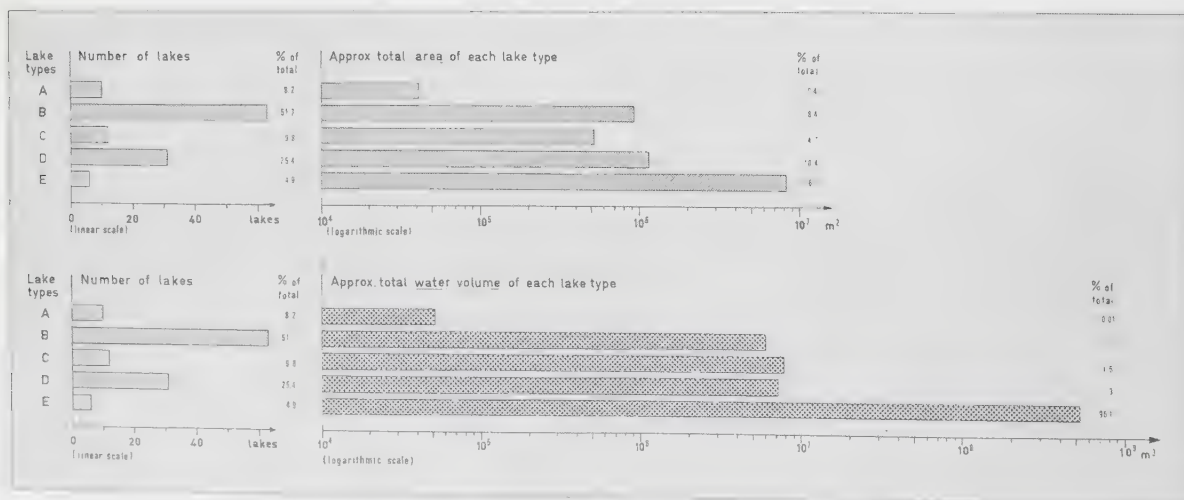


Fig. 6: Statistics according to lake types.

To obtain an estimate of the amount of water dammed up in the lakes of the Expedition Area, the water volume of all the lakes was calculated. As the depth of the ice-dammed lakes could be measured only in a very few cases directly in the field, the water volume of most of the lakes had to be estimated. This was done in a manner similar to that described by Moessner (1963). Although a good knowledge of the lake environments and some indirect indices for the water depth, such as the height of floating and stranded icebergs, helped in the calculations, it has to be stressed that the estimates are in many cases just good guesses. Nevertheless, they give an idea of how much water is dammed up in the different lake classes and the total in all the lakes together. The small lakes can be very important locally but Figure 4 shows that it is the few larger lakes which contain the major part of the water. In Phantom Lake alone there is some 90% of the water volume of all the lakes of the Expedition Area which total about $\frac{1}{2}$ km³. This corresponds to approximately twice the total melt of all the glaciers in the investigation area in a summer like 1960 or some six times that of a summer like 1961 (estimates that were extrapolated from the White Glacier mass budget values of Müller, 1963f, p. 58-59). In contrast, the Aletsch Glacier region has, per unit area, several hundred times less supraglacial and marginal water stored than the Expedition Area.

Thus it is apparent that the Expedition Area with its sub-polar type glaciers has a concentration of ice-dammed lakes comparable to, if not higher than, that of sub-arctic Alaska with its mainly temperate glaciers, and certainly much higher than that of temperate regions such as the Swiss Alps. Flint (1957, p. 174) writes that "lakes formed by blocking of normal drainage by glacier ice are rare today...". This is true for temperate glacier regions at the present time (compared with the situation in the Pleistocene) but it is not correct for the glacierized regions in the Arctic and Sub-Arctic.

b) Types of Ice-dammed Lakes

Water can be dammed up in various ways by glaciers as already discussed systematically by Rabot (1905), Lüttsch (1915), Collet (1925), and more recently by Charlesworth (1957), and Hutchinson (1957). In the literature different names are used for lakes ponded by glacier ice; the most important ones are: "glacial lakes" (Davis, 1882, p. 350; Flint, 1957, p. 142, 143), "glacier lakes" (Charlesworth, 1957, p. 453), "Gletscherseen" (Lüttsch, 1915, p. 3), "glacial reservoirs" (Rabot, 1905), "glacier-margin lakes" (Flint, 1957, p. 174), "lacs d'obturation glaciaire" (Tricart and Cailleux, 1962, p. 341), "ice barrier basins" (Davis, 1882, p. 350), "lacs de barrage glaciaire" (Lliboutry, 1965, p. 665), "Gletscherstauseen" (Drygalski and Machatschek, 1942, p. 251-252; v. Klebelsberg, 1948, p. 382), "Eisstauseen" (Woldstedt, 1954, p. 55), "glacier-dammed lakes" (Liestøl, 1956; Hutchinson, 1957, p. 50; Marcus, 1960, p. 89), "ice-dammed lakes" (Thorarinsson, 1939b; Stone, 1955; 1963b; Hutchinson, 1957, p. 53; etc.). Not all of these names embrace exactly the same group of lakes.

It is not attempted here to introduce a new classification of ice-dammed lakes, but for the convenience of this survey the ice-dammed lakes in the Expedition Area have been differentiated into 5 major types according to site (Fig. 5). The frequency of these types as well as their relative importance as water reservoirs is shown in Figure 6.

Supraglacial lakes (type A). For lakes on the surface of glaciers the following names are used: "superglaciales seen" or "supraglacial lakes" (Lüttsch, 1915, p. 4; Charlesworth, 1957, p. 450), "lacs surglaciaires" (Lliboutry, 1964, p. 370), "surface lakes" (Rabot, 1905, p. 534), "lakes on the surface of glaciers" (Marcus, 1960, p. 89), "surficial lakes" (Bauer, 1961, p. 23), "Eisseen" (Schweizer, 1957, p. 87), "entonnoirs" (Renaud, 1936, although Renaud describes under this name a special type of supraglacial lake); Hutchinson (1957, p. 50) designates them as "type 23a".

In the Expedition Area the total water volume of the supraglacial lakes is insignificant (Fig. 6) as most of them are very small and shallow, often no more than slush pools (Müller, 1962b, p. 306).

Marginal lakes (type B). Lakes ponded between glacier and valley wall, often situated behind a spur or at the mouth of a little ravine, belong to this group. Not included are lakes in tributary valleys, where a substantial part of the lake water comes from the tributary river or glacier; these are classified under types D or E. In a few cases it was rather difficult to decide whether a lake was of type B or D, but these instances were so few as to be of no matter.

Most authors use the term "marginal lake" (Charlesworth, 1957, p. 451), "lac marginal" (Lliboutry, 1965, p. 665), "border lake, lac de bordure" (Rabot, 1905, p. 534), or "Gletscher-Randsee" (Lütschg, 1915, p. 14) in a less restricted way than in this survey, including lakes which are here grouped under C, D or E. Hutchinson's "type 24c" (Hutchinson, 1957, p. 53) embraces lakes grouped under type E below.

The marginal lake is the most frequently occurring type of ice-dammed lake in the Expedition Area (Fig. 6) as well as in the rest of Axel Heiberg Island. Most of them are of small or medium size.

Lakes at the junction of converging glaciers or glacier arms or below nunataks (type C). In the literature they are described under different names, e.g. "lacs au point de confluence de deux glaciers" (Collet, 1925, p. 94), "nunatak lakes", "lee-lakes", 3rd variety of "marginal lakes" (Charlesworth, 1957, p. 450-451), "Typ Gorner" (Schweizer, 1957, p. 86), "type 24c" (Hutchinson, 1957, p. 53). This is not a new type as Ricker (1962) suggests; such lakes have been mentioned at an early date (de Saussure, 1804, p. 396; Agassiz, 1841, p. 204; Davis, 1882, p. 352; Russell, 1893, p. 234; Rabot, 1905, p. 538; etc.).

Some 10% of all the lakes in the Expedition Area belong to this group (Fig. 6). Most of them are fairly small, but one of them, the Between Lake, is larger than 400 000 m² and contains some 7 500 000 m³ of water.

Lakes in tributary stream valleys (type D). A main valley glacier may dam up water in a tributary valley which is usually relatively small and not occupied by a glacier. A good part of the lake water comes from the tributary river. It is not always clear whether a lake belongs to this group or is of the B or E type.

Davis (1882, p. 350) describes these lakes under his first "subspecies of ice barrier lakes"; Rabot (1905, p. 534) classes them as a variety of "border lakes", Lütschg (1915, p. 14) as "Gletscher-Randseen"; Charlesworth (1957, p. 451) describes them under his "marginal lakes"; Schweizer (1957, p. 86) classifies them as "Typ Märjelen" and Hutchinson (1957, p. 51) as "type 24a". Such lakes have already been mentioned in very early writings on glacierized regions (Fabricius, 1788, reprint in English 1955, p. 409; Agassiz, 1841, p. 203).

About one fourth of all the lakes of the Expedition Area belong to this important group. More water is stored in them than in all the marginal and supraglacial lakes together (Fig. 6).

Lakes in tributary glacier valleys (type E). Marcus (1960, p. 90) defines them as lakes "formed by the retreat of a tributary arm from the trunk glacier". Many authors make no separate mention of such lakes, but some do, particularly Rabot (1905, p. 534), George (1947, p. 81-82), Stone (1955, 1963b), Liestøl (1956, p. 132-135), and Marcus (1960, p. 90). Hutchinson (1957, p. 53) describes a similar lake and calls it "a rather special type of ice-dammed lake". There is no such lake at present in the Aletsch Glacier region, although one might anticipate the development of one at the junction of the Mittelaletsch Glacier with Aletsch Glacier. Observations from British Columbia and Alaska (Stone, 1955, 1963b; Marcus, 1960) from Norway (Liestøl, 1956) and from Axel Heiberg Island (Ricker, 1962; Maag 1963, p. 155) show that it is an important lake type in those areas.

Although not very numerous in the Expedition Area, these lakes are very important as water reservoirs. The three largest lakes belong to this type; they contain some 95% of the lake water of the area, Phantom Lake alone having some 90% (Figs 6 and 10).

Miscellaneous lakes. A lake dammed up in a transverse valley between two parallel valley glaciers or ice lobes, such as that described by Stone (1955, p. 77), does not occur in the Expedition Area but exists in other parts of Axel Heiberg Island. Likewise lakes dammed up in the main valley by a tributary glacier ("type 24b" of Hutchinson, 1957, p. 50; "Typ Mattmark und Rofen" of Schweizer, 1957, p. 84-86) exist on Axel Heiberg Island, but not in the study area although one such lake might be in the process of development behind the Phantom Glacier tongue which protrudes into the Phantom Lake valley and already bars part of the meltwater of the Transit Glacier (Figs 32, 34).

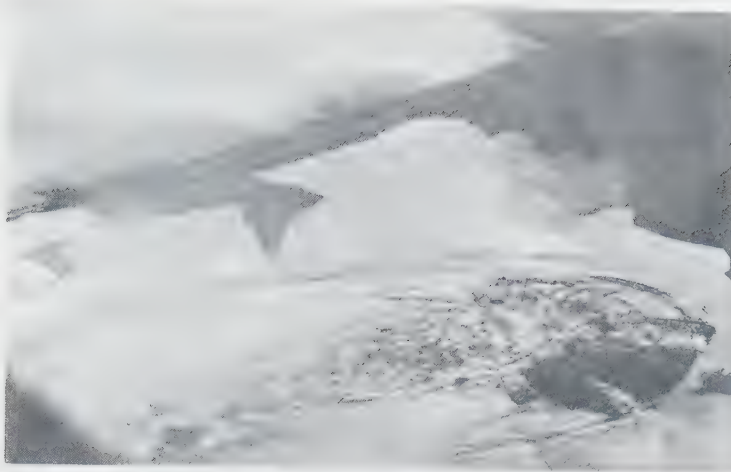


Fig. 7: Marginal area of Steacie Ice Cap. Large radial and concentric cracks, (diameter: several 100 m); water in the centre and at margin of cracked area (oblique air photo, summer 1960)

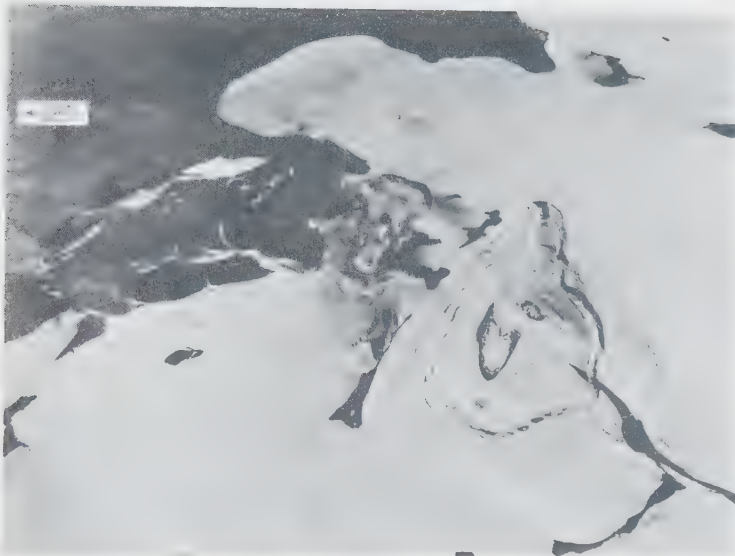


Fig. 8: Large depression (diameter: several 100 m), water in centre, situated at about 80°30'N, 92°50'W (air trimetrogon photo T412 R-146, 16 July 1950).

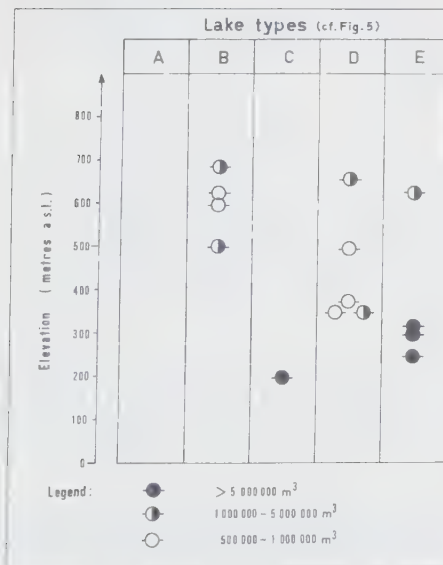


Fig. 9: Largest lakes in Expedition Area, summer 1960.

Lakes dammed up by moraines (Hutchinson's "types 25, 30, 32 and 33", 1957, p. 56, 82, 90) were only counted as ice-dammed lakes if it was obvious that the ice and not the moraine was ponding the water. Such lakes are rare in the Expedition Area. At some places there are shallow ponds on the surface of ice-cored moraines (Fig. 85) but they have not been included in this survey.

"Type 24e" of Hutchinson (1957, p. 54) - "proglacial lake at the edge of a waning ice sheet" - is not represented in this mountainous area which contains only part of a relatively small ice field (Hidden Icefield), the little lakes at the edge of which are grouped under "marginal lakes" and "lakes in tributary valleys".

Also at present in the Expedition Area there are no ice-dammed lakes ponded by glaciers that block off the head of a fiord (Hattersley-Smith and Serson, 1964, p. 108-111), nor are there lakes dammed by an ice shelf.

No sub- and englacial lakes (Hutchinson's "type 23b", 1957, p. 50) were encountered, and they probably do not exist in the Expedition Area because of sub-zero ice temperatures.

Three interesting lakes discovered outside the Expedition Area merit closer inspection on the ground. 1) In a depression on the Steacie Ice Cap radial and concentric cracks separate large ice blocks; this was photographed by F. Müller from the air (Fig. 7). It might be a similar phenomenon to that observed in Spitsbergen by Vivian (1965, p. 379). 2) On air photographs (T412) R-146/147, A 16 754 - 130/131, A 16 838 - 100/101) a large depression situated at about 80° 30'N, 92° 50'W was seen (Fig. 8) which seems to be a drained supraglacial lake basin lined with the collapsed former ice cover. 3) A lake situated at about 81°N, 93° 15'W (air photographs A 16 836 - 88-90, A 16 754 - 141/142), at the junction of two tributary glaciers which are apparently thinning quite rapidly, is dammed up by a glacier which appears to have been once the joint tongue of the two tributary glaciers but which is now cut off from the main part of its former catchment area. This lake does not properly belong to the group of lakes between converging glacier arms (type C) nor to the lakes forming in tributary glacier valleys (type E).

c) Lake Size and Lake Type

Figure 9 showing the largest lakes and their type and Figure 6 comparing the number of lakes and the total water volume of each group demonstrate clearly that the average lake size differs much from group to group. The largest lakes are found among the lakes in lateral valleys which were formed when a tributary glacier retreated from the trunk glacier (type E). There are also some relatively large ones in group D. Small lakes occur most frequently in groups A, B and C, but very rarely in group E.

So in the Expedition Area (and this applies also to Axel Heiberg Island as a whole) small lakes "prefer" different sites with respect to the glacier than do large ones; this is contrary to observations in Alaska (Stone, 1963b, p. 338).

d) Elevation and Orientation of Ice-dammed Lakes

In 1960, in the study area, ice-dammed lakes occurred from approximately 120 m a.s.l., that is some 50 m above the glacier tongues, up to approximately 1100 m a.s.l. (Figs 10, 11), the uppermost part of the ablation zone. In 1960 and in 1962 supraglacial lakes as high as 1300 m a.s.l. were observed on the McGill Ice Cap (Müller, 1963c, p. 15). But the majority of the lakes lie within a smaller range, 90% between 250 m and 930 m, 70% between 350 m and 780 m, 50% between 410 m and 720 m a.s.l.

Summing up the water stored at the different elevations (Fig. 10), it is evident that the lake water is concentrated in a relatively narrow zone: 96% of the water is ponded in lakes whose surfaces lie between 340 and 500 m a.s.l., that is in the lower part of the middle zone of the ablation area.

It appears that the average elevation of the small lakes is somewhat higher than that of the large ones, which differs from observations in Alaska (Stone, 1963b, p. 338). All surfaces of

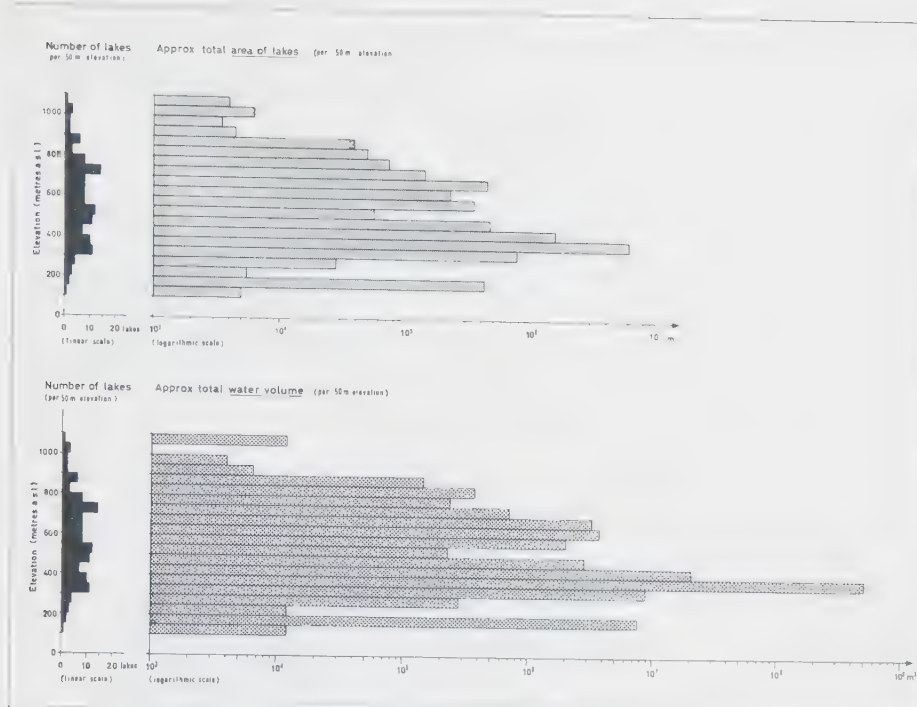


Fig. 10:
Total number, area and
volume of ice-dammed
lakes by 50 m elevation
intervals, summer 1960.

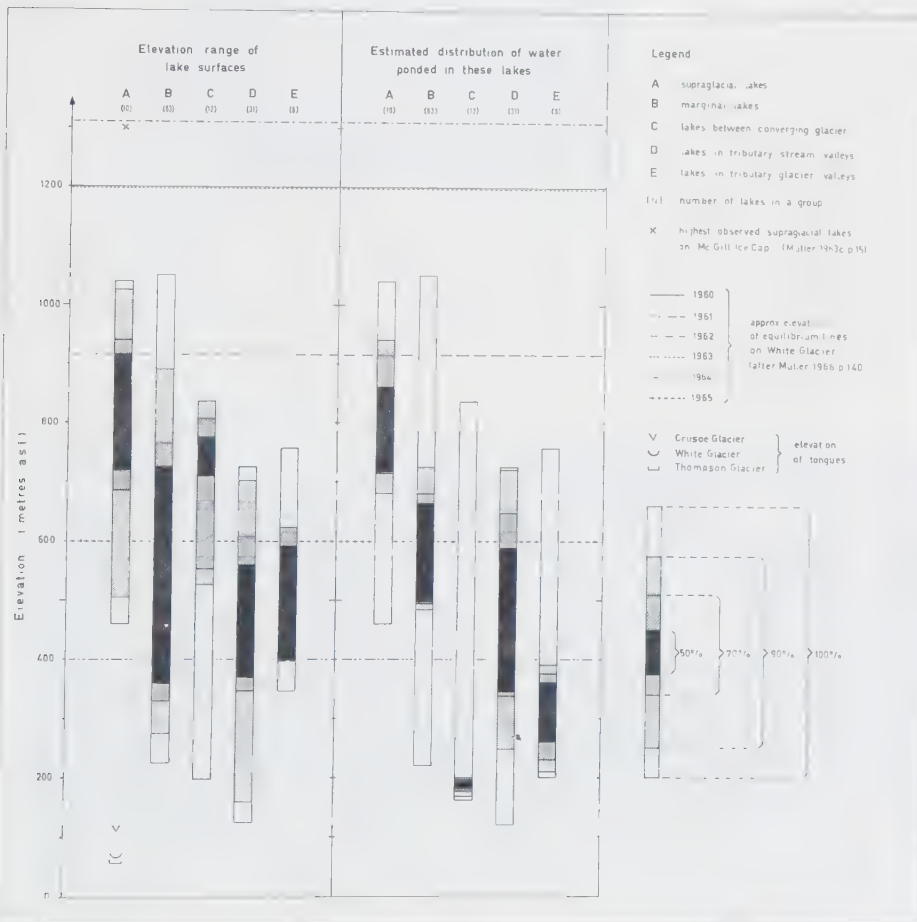


Fig. 11: Elevation range of different types of ice-dammed lakes, summer 1960.

the lakes larger than 100 000 m² lie below 700 m and 3 of the 4 largest lakes (over 400 000 m²) are found in the narrow zone between 347 m and 412 m a.s.l. (Fig. 9).

Figure 11 gives the preferred elevations of the five lake types.

The orientation of all the lakes of the Expedition Area that were included in this survey is shown in Figure 12. No statistically significant difference in number of lakes orientated towards E or W or towards N or S, i.e. no particular preference for any compass orientation, is evident from the diagram. That the number of lakes orientated E or W is larger than those facing N or S can be explained by the fact that the main glaciers run N - S. Checks in other glacier valleys of Axel Heiberg Island corroborate the conclusion, also drawn by Stone (1963b, p. 338) for Alaska, that exposition seems to have no significant influence on the location of ice-dammed lakes although the number of investigated lakes is still too small to allow a final statement.

The main reason for this is probably the northerly latitude where the differences in insolation and meltwater supply between day and night are not as pronounced as in more southerly regions, especially during June and July which are the main replenishing months for the lakes.

Even if there were - despite the northerly latitude - a slight dependency of lake site on exposition, it probably would not manifest itself because of the stronger influence of local environmental conditions such as the colour of adjacent rocks, etc., and particularly the geological structure and valley direction.

e) Ice Cover of Lakes

At the end of July 1959 about 1/6 of all the ice-dammed lakes were still covered with more than 50% of lake ice and about 1/4 with more than 25%. Such remnants of winter ice might give some indirect evidence of which lakes did not empty completely the previous summer and continued partially or totally filled through the winter. The amount of ice cover remaining at the end of July depends on the character of the summer weather (e.g. at the end of July 1961 greater areas of the lakes were still ice covered than in 1960), the amount of winter and spring snow and probably also on the elevation of the lakes, although no definite correlation could be seen. A more important influence may be exerted by local circumstances such as extreme shadow due to deep entrenchment (Møller, 1960, p. 188), water temperature (dependent on type of influents, relative length of damming ice front, number of icebergs, local winds, etc.). A check of air photographs outside the Expedition Area showed that in the northern part of the island comparatively more ice-dammed lakes were still fully ice-covered than in the study area, although there is only a difference in latitude of about 1°.

f) Type of Drainage

Almost all the ice-dammed lakes observed in the field or identified from the air or on air photos show signs of marginal or supraglacial drainage in the form of outlet channels at the margin between rock and glacier, on the glacier surface, or over a col. This differs from the drainage of the majority of the Alaskan ice-dammed lakes (Stone, 1963b, p. 340). A further discussion of drainage mechanisms is given in 2b and 2c of this chapter.

g) Geology and Lake Site

If one compares the distribution of lakes with the geological map (Fricker, 1962) it is obvious that the location of many ice-dammed lakes is dependent on certain geological conditions as, for example, the presence of dykes, sills, fault lines, folds, or change in the geological formation that favour the ponding of water at that particular place. Such a dependency may be more pronounced in arctic than in temperate glacier regions as impermeable rock and soil (due to permafrost conditions) and impermeable glacier ice (due to sub-zero ice temperatures) prevent the water from detouring, for instance, a structural barrier through the soil, rock or glacier.

2) Filling and Emptying of Ice-dammed Lakes

a) Methods of Investigation

To record lake level fluctuations staff-gauges and chain-gauges were used (Fig. 13b). Often

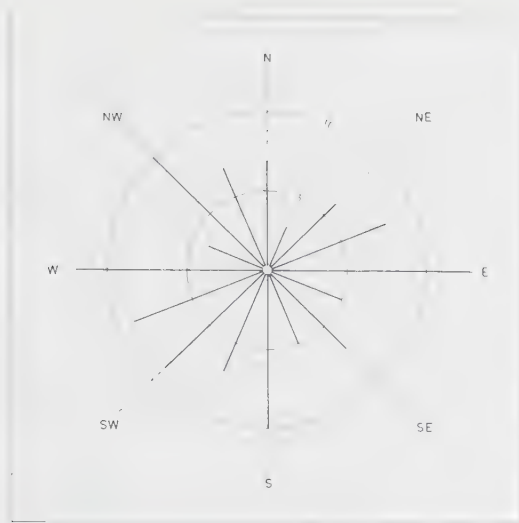


Fig. 12: Orientation of ice-dammed lakes in Expedition Area, summer 1960.



Fig. 13a: Between Lake, dammed by the coalescing tongues of Thompson and White glaciers 17 hours after start of overflow; distance from photo standpoint to overflow point some 1100 m (24 July 1963, 0930).



Fig. 13b: Between Lake, 76 hours after start of overflow (26 July 1963, 2115)

Table 4: Between Lake, filling and emptying 1959-1965

max. water level (1963) approx. 195 m a.s.l.		max. lake surface area (1963) approx. 410 000 m ² (1)		- see Fig. 14	
" " depth (1963) " 35 m		" water volume (1963) " 7 500 000 m ³ (1)		" "	

(1) estimated volume (Maag, 1963, p. 154) was too low,

(2) 28 July 1959: lake completely empty (air photo A16755-116/117),

(3) see remnants of old channels in air photos,

(4) Ricker's emptying time of 2 days (Ricker, 1962) for whole lake does not agree with photos & statements of others,

(5) see also air photo 10B-229/230, 2 Aug. 1960,

(6) personal communication & photos of C.S.L. Ommanney and H. Weiss,

(7) personal communication & " " C.S.L. Ommanney,

(8) 29 July 1965, lake still emptying (personal communication and photos of C.S.L. Ommanney).

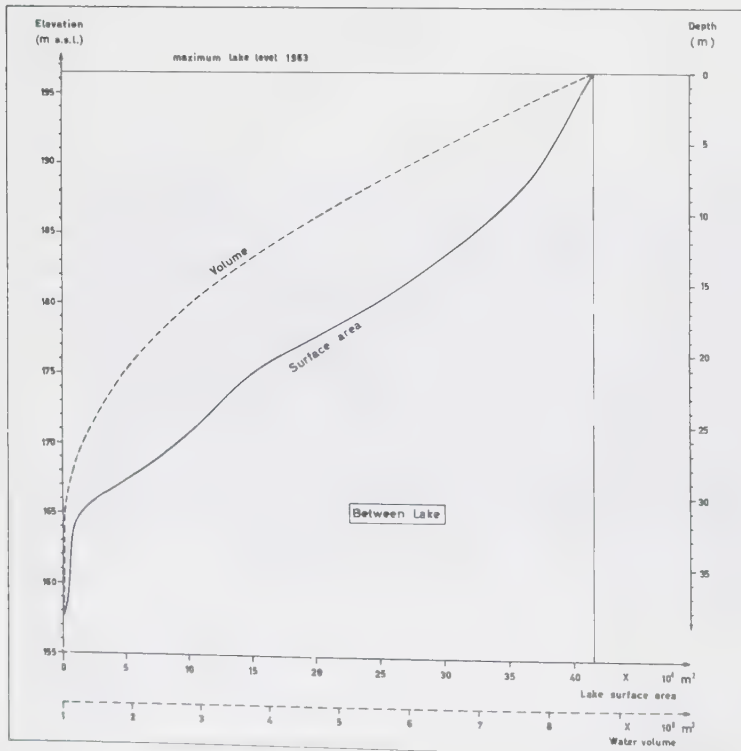


Fig. 14: Bathygraphical curves of Between Lake (approximate water volume and surface area at various filling stages).

a set of gauges had to be installed so that the whole range of possible level fluctuations was covered. Maximum levels were marked with paint and cairns. Repeated photographs from fixed points and from airplanes were valuable aids in recording the stages of filling and emptying of lake basins. In addition, shorelines, stranded ice pieces, mud cover, etc., were often useful indicators of lake fluctuations and stagnations.

As in many cases it was very difficult or even impossible to measure the exact volume of the outflowing water, a continuous record of the lake level change was especially important. Although the different gauges were read as frequently as possible, many important level fluctuations were missed. Some lakes could only be visited sporadically. To obtain a continuous record automatic level gauges (limnigraphs) would have to be installed. The pneumatic types would be most useful as the shore conditions often make it difficult to install level recorders working on the float principle.

b) Observations from 4 Lakes

Between Lake: This triangular shaped lake dammed up between the Thompson Glacier and White Glacier tongues (Fig. 13a,b) is the most investigated lake of the Expedition Area. Surface area and volume at different lake levels are shown in Figure 14. According to a preliminary estimate (Beschel, 1961a, p. 192), the present lake started to form about 50 years ago.

Table 4 gives main observations regarding filling and emptying during the period 1959 to 1965. The filling speed and the onset of drainage differ from year to year, depending to a large degree on the meltwater supply. "Normally" the Between Lake drains in the middle of the summer. But when the summer weather is as poor as in 1964 there is not sufficient meltwater to fill the basin and the lake remains filled over the winter; this seems to be the exception. Rain storms that caused marked increases in river discharge (Adams, 1966, p. 63, 67-68; and 3b of Chapter C) resulted in only a minor rise in lake level (Fig. 15) which shows how such large lake basins have a "dampening" effect in times of flood.

The highest water levels attained in the years since 1959 did not differ very much from year to year; from 1959 to 1964 they remained a little less high than in earlier years as manifested by old shorelines (Fig. 16). In 1965, the lake was apparently dammed up to a maximum height (personal communication of C.S.L. Ommanney). This might be explained in the following way: there was no cutting of an outflow channel in 1964; the forward-pushing Thompson Glacier (Müller, 1963g, p. 75-76) could therefore exert its pressure on the White Glacier tongue without first having to close such a channel, so that during the summer of 1964 and the winter of 1964/65 the overflow area was probably pressed up higher than usual. In addition, pieces of the broken up winter ice cover and some icebergs had piled up at the outlet helping to dam up the lake water to a maximum height.

From the observations on Between Lake it is evident that since 1959 - and very likely also prior to that date - it always drained in the same way, i.e. by overflowing the damming ice at the lowest point, which is approximately at the junction of the two glacier tongues, cutting (by melting and by mechanical erosion) a deep channel into the ice surface (Fig. 17a, b). The differing dynamics of the two glaciers may be the main reason for the lower part of the channel being cut entirely across the surface of the White Glacier tongue and not following the line of junction of the two glaciers, as might be expected (Fig. 20).

The speed of drainage varies from year to year (Fig. 18) depending mainly on the availability of open channel stretches (Fig. 19). The observations showed that, despite the pressure exerted by the glaciers, the channels formed in previous summers do not usually close completely and are often re-used during subsequent emptyings; over short stretches some channels just close at the surface but remain open underneath (Figs 19, 20). On the other hand, the overflow of 1962 proved that one drainage is sufficient to erode a whole new channel. If the outlet channel collapses during the drainage, as was observed in 1961 and again in 1963, or if it is plugged with icebergs and ice pieces, as was the case especially in 1965, this leads to a slowing down of the outflow or even to minor temporary pondings; however, the water soon finds a way through the ice plug (Fig. 18). The temperature of the lake water also seems to have a decisive influence on the speed of the outflow. In 1965, for example, due to the large amount of lake ice pieces and icebergs, the water must have been colder than in previous years which might have been one of the causes for the slow drainage of that year.



Fig. 15: Discharge of Ermine River and water level of Between Lake during rain storm, July 1961.

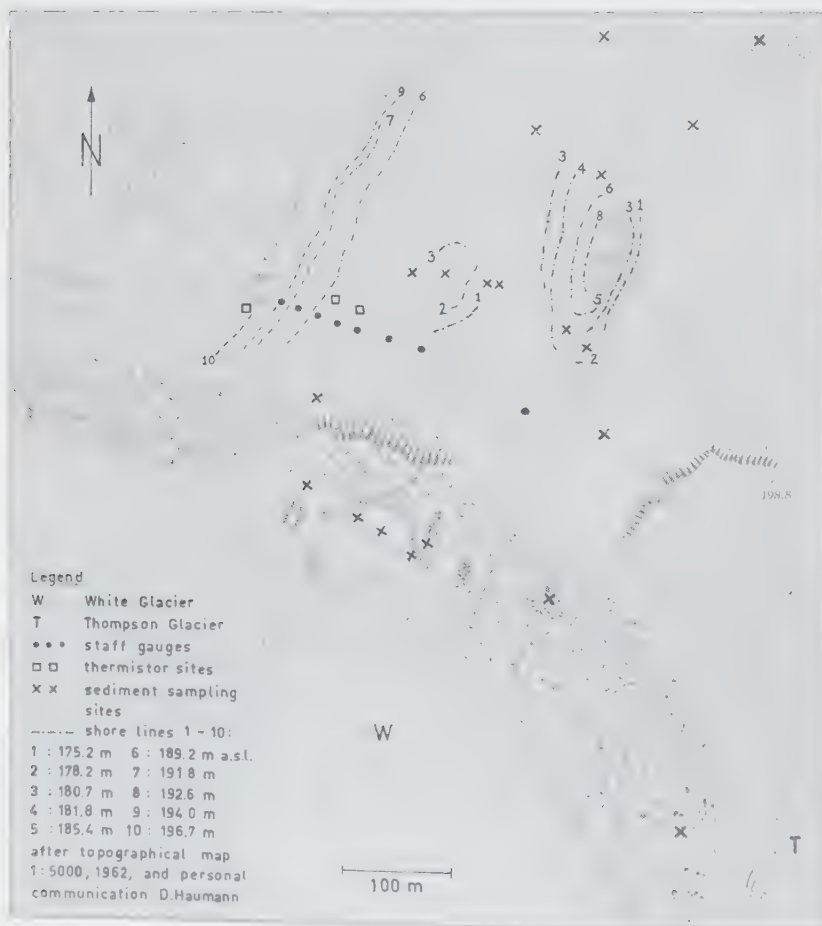


Fig. 16: Shorelines, gauges, thermistor sites, and sediment sampling places in Between Lake.



Fig. 17a: Outlet channel of Between Lake about 12 hours after beginning of overflow: cairn on right 60 cm high (24 July 1963, 0500).

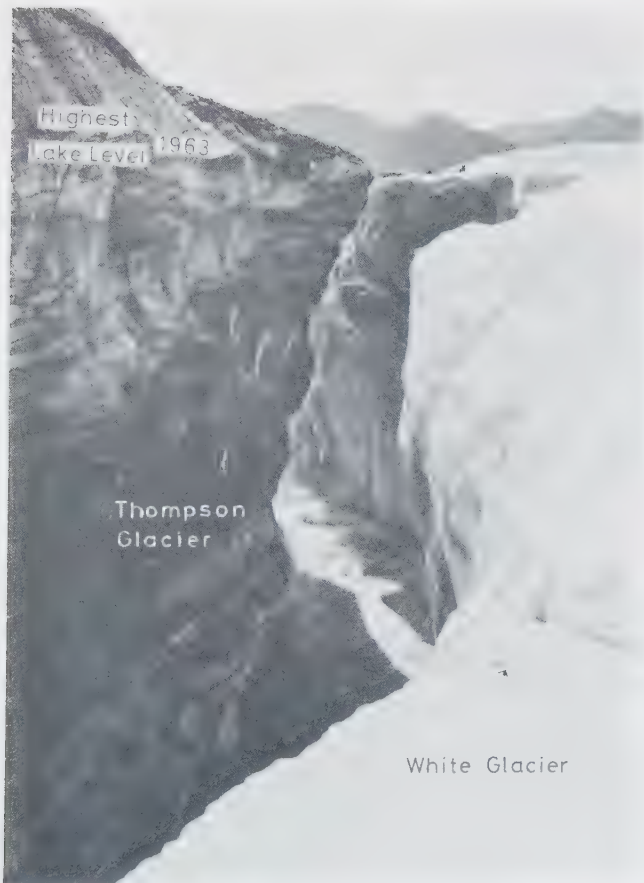


Fig. 17b: Outlet channel of Between Lake about 40 hours after beginning of overflow, approximately 14 m deep (25 July 1963, 0900).

Fig. 18a: Water level of Between Lake during drainage, summer 1961 and 1963.

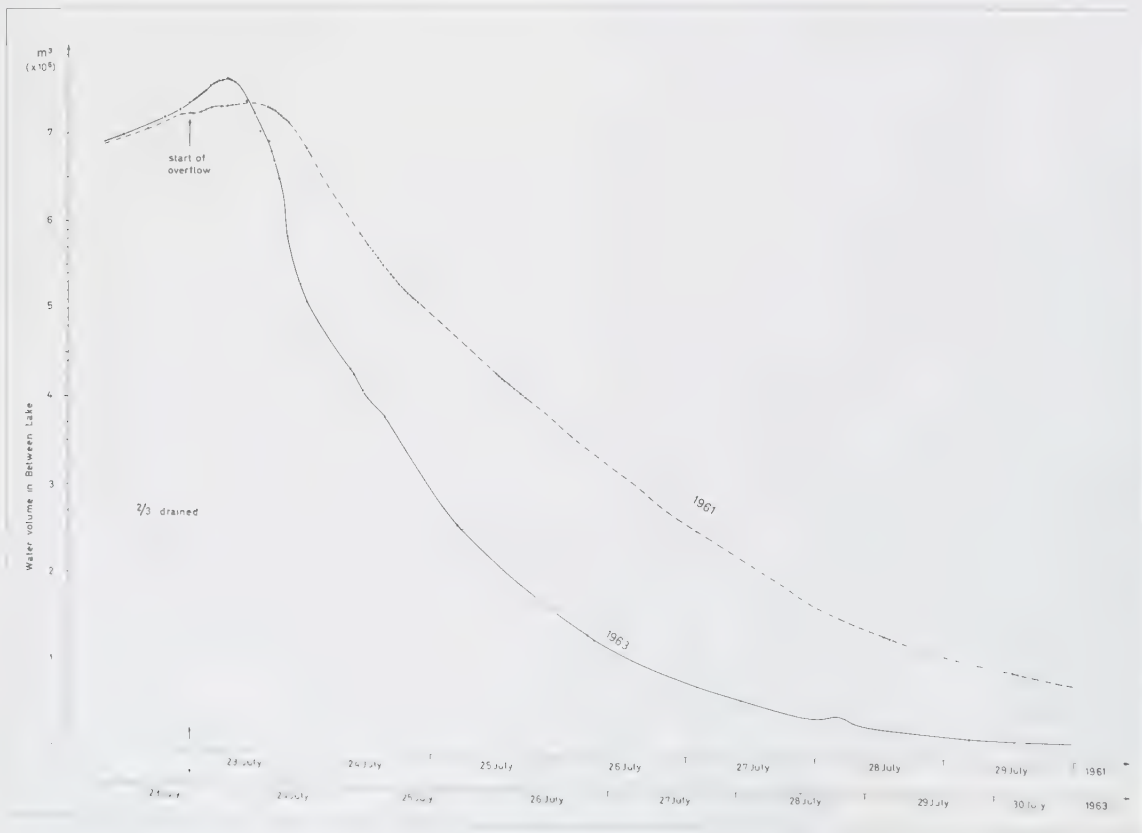
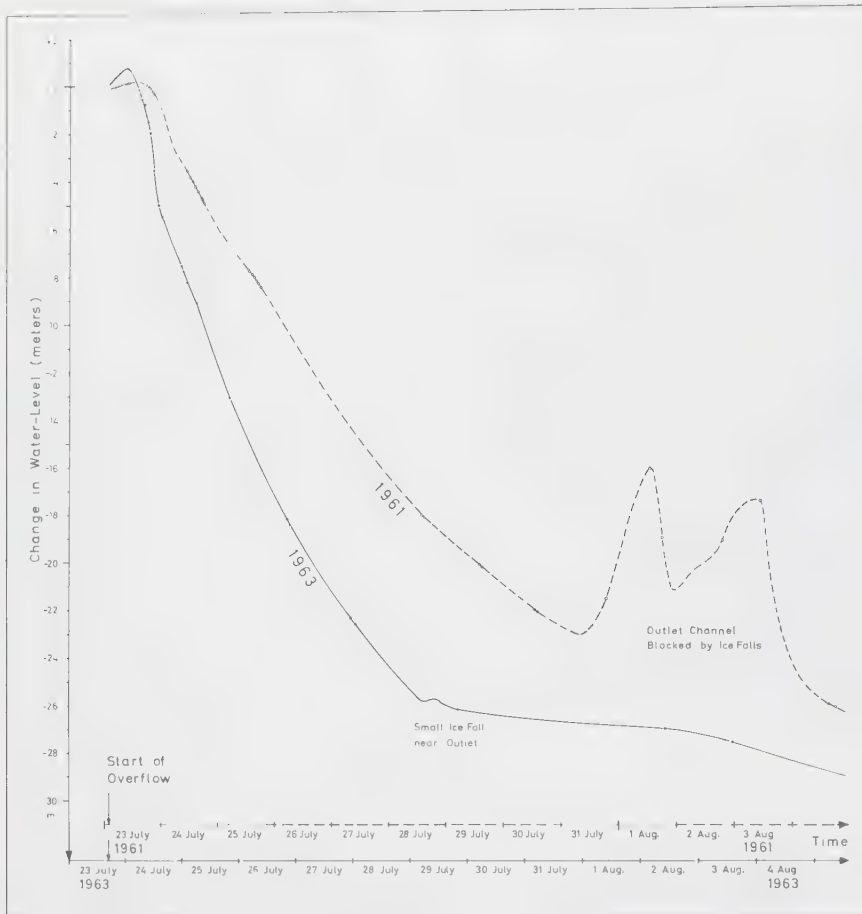


Fig. 18b: Water volume of Between Lake during drainage, summer 1961 and 1963.

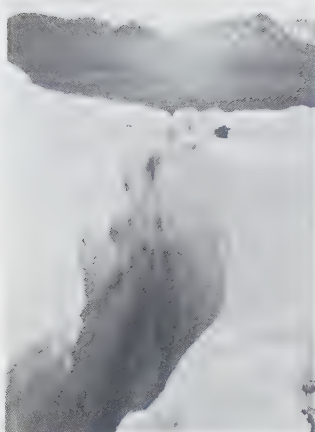


Fig. 19: Channel cut into White Glacier tongue by outflowing Between Lake water (30 June 1963).

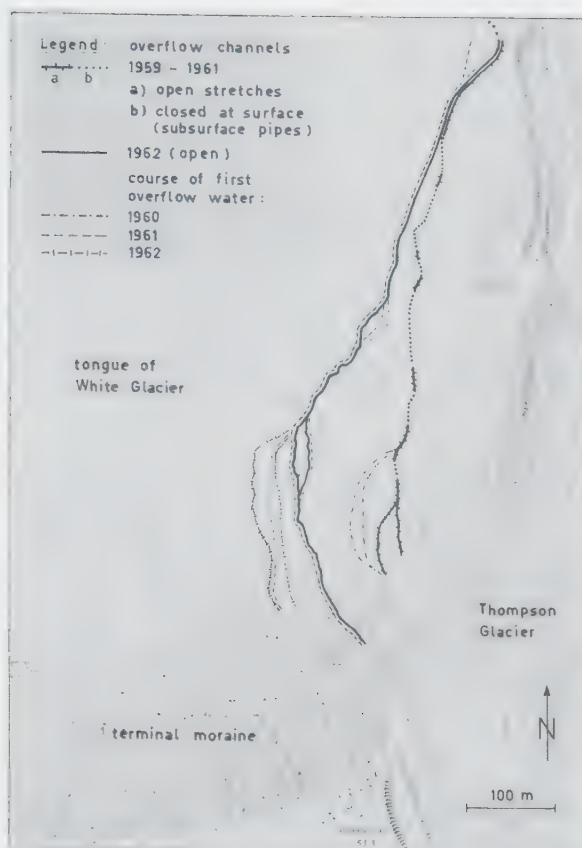


Fig. 20b: 1959-1962 channels on White Glacier tongue.



Fig. 20a: Tongue of White Glacier with Between Lake overflow channels (oblique air photo, summer 1962).

Crusoe-Baby Lake: This kidney-shaped lake is dammed up by the left margin of the Crusoe Glacier at the mouth of a tributary river valley (Figs 21, 22). The tributary river brings water mainly from a small niche glacier, Baby Glacier; in addition, meltwater flowing marginally along the Crusoe Glacier collects in this lake basin. The lake site is influenced by structure; the place where the water is dammed up by a rock spur and the glacier ice coincides with the borderline between two geologic formations (Fricker, 1962, Preliminary Geological Map).

In Table 5 some data on the emptying of the Crusoe-Baby Lake are assembled. The maximum lake level did not differ much from year to year; in 1962 and 1963 it was a few centimetres higher than in 1961. Since observations started the lake emptied annually, usually in the first half of July, with the exception of 1964 when due to poor weather conditions the overflow began several weeks later. The lake drained always in the same way, namely by overflowing marginally and cutting a channel between the ice margin and the slope (Fig. 23). The upper part of this marginal outlet channel was observed to collapse after several hours or days (Figs 24, 42). After such a collapse the lake might be dammed up again for a short time but in general the water soon finds a way to the lower part of the buried channel and the drainage continues sub-marginally.

In comparison with other lakes Crusoe-Baby Lake emptied relatively slowly (Fig. 25). The main reasons for this might be due to the various collapses of the outlet channel, or the low melting capacity of the lake water which is relatively cold because of the shadowy site and the high percentage of glacial meltwater flowing in. In addition, the normally abundant water supply during the height of the ablation season was considerably increased in 1961 and again in 1963 by the contemporaneous drainage of other ice-dammed lakes situated higher up. On the other hand, the existence of old marginal and sub-marginal channel stretches which did not completely close (Fig. 74) might sometimes help to speed up the drainage process. In 1961, at the beginning of the outflow, the lake level fell a little but started to rise again for several hours despite the continuing drainage (Fig. 25). This was due to the additional water supply from heavy rainfalls preventing the outflow from keeping up with the inflow although the outlet channel was deepened relatively fast.

Ice Cave Lake: Some 500 m down-glacier from the Ermine Lake, the Ermine River disappears in summer under the left margin of the White Glacier into a near-marginal tunnel, the Ice Cave (Fig. 26). In winter this passage is usually blocked by ice falls and snow as observed in the spring of 1961 by Professor F. Müller and the author who penetrated this cave to a depth of about 150 m. This barricade dams up the first waters of the Ermine River to form a lake, the Ice Cave Lake. Surface area, depth and volume of this lake are given in Table 6 and Figure 27.

The course of the Ermine River between Ermine Lake and the Ice Cave coincides approximately with a faultline (Fricker, 1962, Preliminary Geological Map) which seems to run under the glacier just about where the river disappears into the Ice Cave or forms the Ice Cave Lake.

Probably the Ice Cave was once a normal marginal river bed which became sub-marginal as the White Glacier grew thicker (Fig. 28) and overrode its own lateral moraine and marginal river (Maag, 1963, p. 158).

In Table 6 some data on the filling and emptying of the Ice Cave Lake are summarized. During the period of observations the lake formed annually at the same site and usually emptied rather early in the melting season. The maximum lake level varied from year to year depending mainly on the thickness of the damming glacier margin. In 1963, for example, the maximum lake level was some 3 m lower than in 1961 (Fig. 29), confirming the rapid thinning of the White Glacier in this section (Müller, 1963g, p. 72-74). Although this is a relatively small lake, the observations might be of some interest because of the special way it drained. In 1961, the lake was dammed up three times consecutively (Fig. 30). The first time the overflow commenced marginally and rather slowly but after a few hours it began to drain through the Ice Cave and more rapidly. The pressure of the dammed up water and its melting capacity probably opened the tunnel. The second time also the lake started to overflow marginally and then after a very short time changed to drainage through the Ice Cave. The third time - after just a partial damming up - drainage took place entirely through the Ice Cave. Afterwards, during the whole summer, the river flowed freely through this passage. Two years later, in 1963, the lake drained somewhat differently. As previously it started by overflowing between the ice and the slope, but instead of using the tunnel it continued to drain marginally, cutting a deep marginal and sub-marginal channel (Fig. 29), similar to the Crusoe-Baby Lake



Fig. 21: Crusoe-Baby Lake dammed up by Crusoe Glacier, 22 July 1961. Dotted line shows highest water level of 1961 (5 July).

Table 5: Crusoe-Baby Lake, Drainage 1959-1965

maximum water level (1961) : approximately 250 m a.s.l. " " depth (1961) : " 20 m " lake surface area (1961) : " 20 000 m ² " water volume (1961) : " 260 000 m ³ " discharge (1961) : " 3 m ³ /sec.			
Year	Emptying Dates	Kind of Drainage	Remarks
1959	Complete empty prior to 28 July ⁽²⁾	Probably marginal and then sub-marginal as 1961	Lake was empty at least one week earlier than in 1961
1960	Beginning of July ⁽³⁾ Complete empty prior to 2 August ⁽⁴⁾	Probably as 1961	
1961	5 Jul: start of overflow after warm rain 9 Jul: first collapse of undercut ice-margin After some 12 days about 2/3 of lake were drained 5 Aug: lake basin completely empty	Marginal overflow between glacier and slope, becoming sub-marginal by undercutting the glacier edge (Fig. 23)	Further downstream old channel sections were re-used Drainage slower than at Between Lake
1962 ⁽⁵⁾		As 1961	Re-use of old channel sections as in 1961
1963	About 1 Jul: start of overflow 16 Aug: still emptying	As 1961 but higher damming up when the outlet channel collapsed the first time	As 1961
1964 ⁽⁶⁾	Start at least 3-4 weeks later than 1961 16 Aug: still emptying	Similar to 1961	Outflow probably slower than in 1961 as water was colder
1965 ⁽⁶⁾	Empty prior to 20 August	Similar to 1961	

(1) Estimate of area and volume in Preliminary Report 1961-62 (Maag, p. 151) too low.

(2) See aerial photos of 28 July 1959 (A 16 836 - 31/32)

(3) Personal communication F. Müller

(4) See aerial photos of 2 August 1960 (054 -056; 106/107)

(5) Personal communication and photos C.S.L. Ommanney and H. Weiss

(6) Personal communication and photos C.S.L. Ommanney

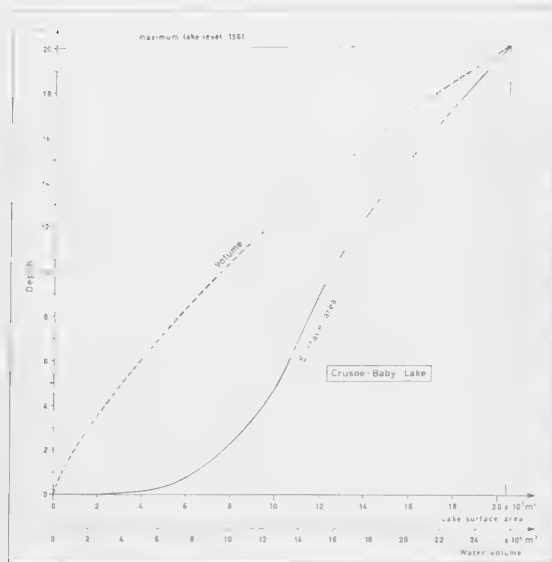


Fig. 22: Bathymographical curves of Crusoe-Baby Lake (approximate water volume and surface area at various filling stages).



Fig. 24: Outlet channel of Crusoe-Baby Lake with evidence of two collapses of ice wall (1,2). Shorelines and kolks on ice cliff show lake levels. Dotted line indicates maximum lake level, approximately 17 m above present level (16 July 1963, 2200).

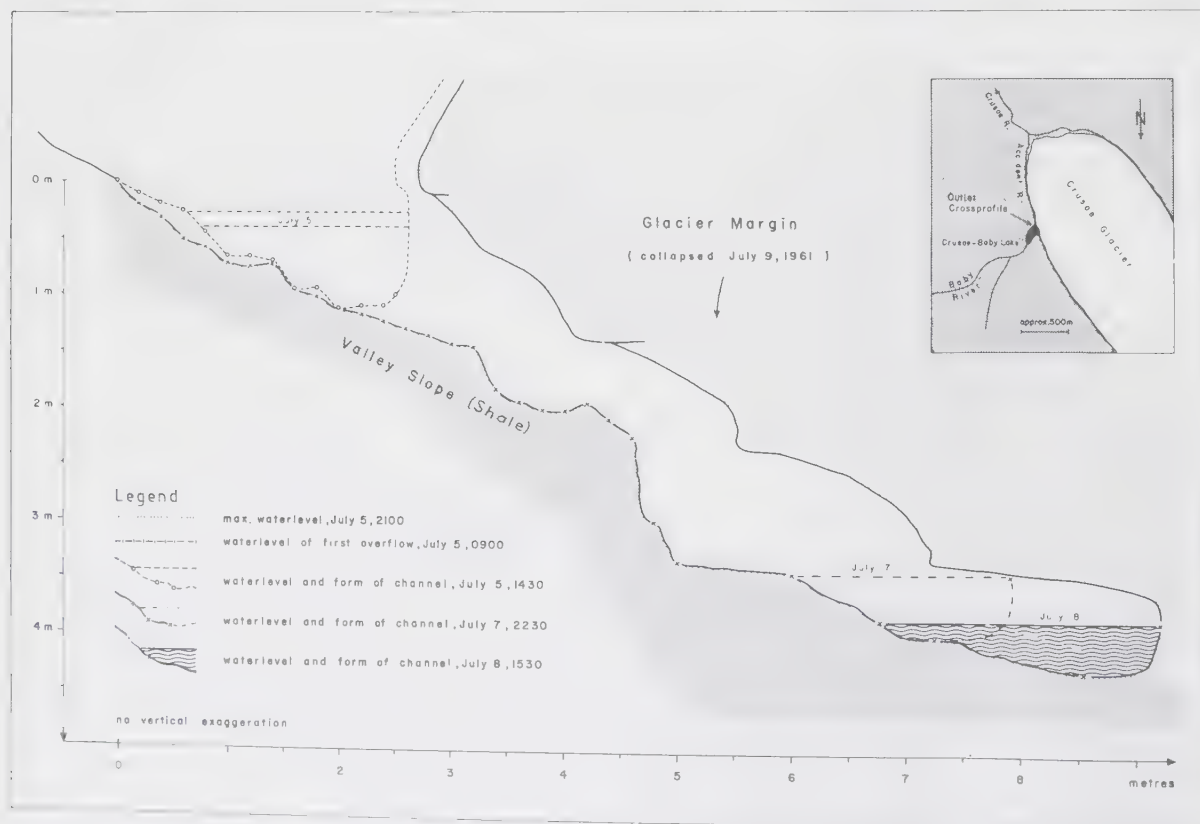


Fig. 23: Cross-section of outlet channel of Crusoe-Baby Lake, progressive change of marginal channel during first days of lake drainage, 5-8 July 1961.

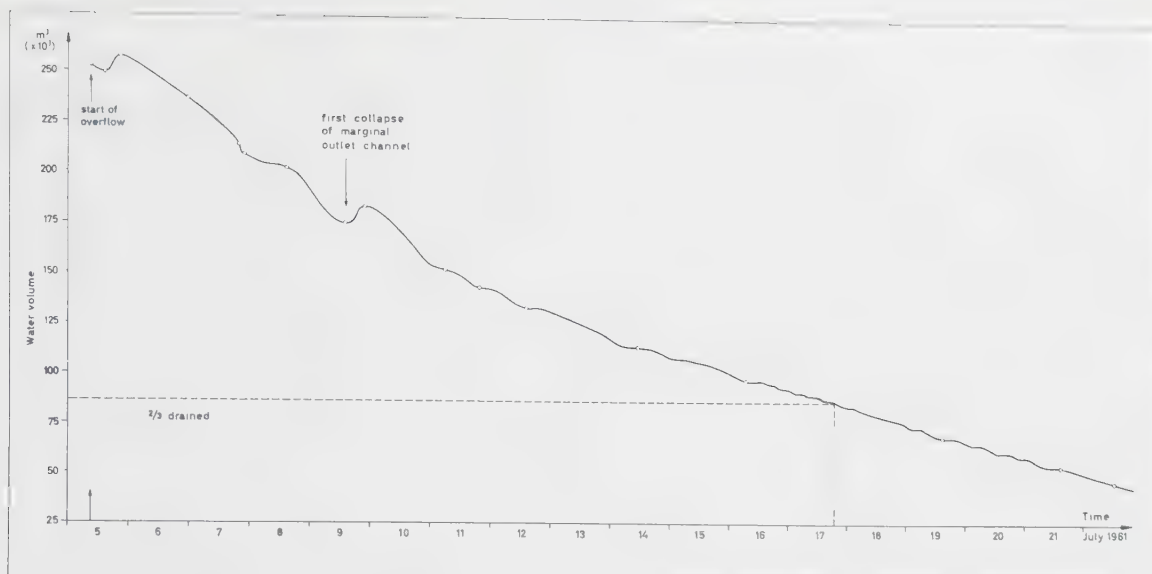


Fig. 25: Water volume of Crusoe-Baby Lake during drainage, summer 1961.



Fig. 26a: Ice Cave Lake during drainage, 24 June 1963, 1800. Dotted line shows highest water level 1963.



Fig. 26b: Ermine River disappearing into Ice Cave (x). Rod for scale approximately 4 m, (17 Aug. 1961).

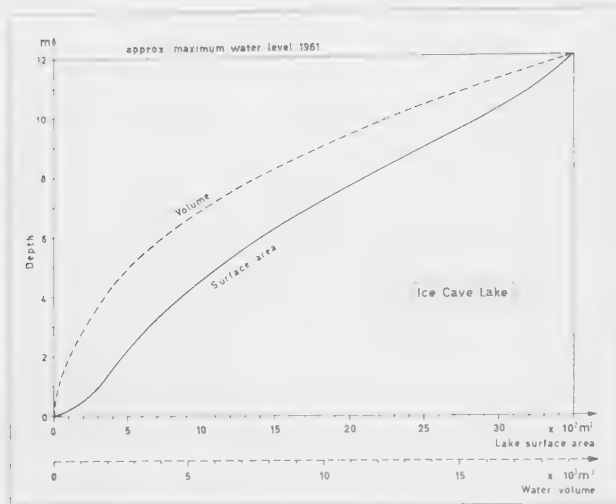


Fig. 27: Bathymographical curves of Ice Cave Lake.

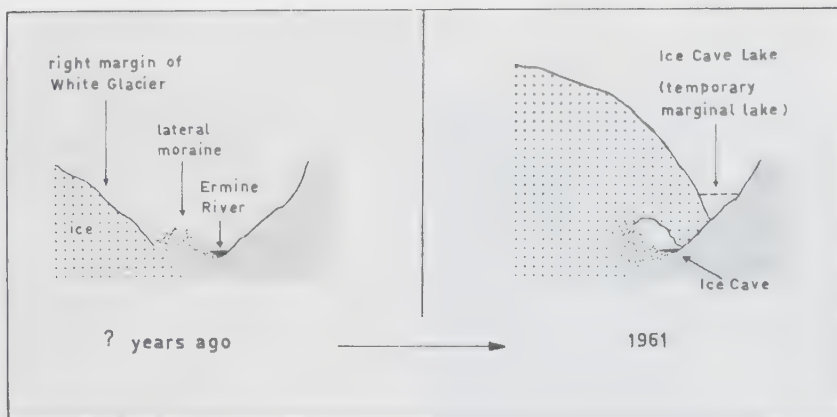


Fig. 28: Genesis of Ice Cave, hypothetical cross-sections (not to scale).



Fig. 29: Margin of White Glacier with outlet channel of Ice Cave Lake, 24 June 1963; highest water level of 1963 some 3 m lower than in 1961.

Table 6: Ice Cave Lake 1960-1964

max. lake level (1961): approx. 318 m a.s.l.		max. surface area (1961): approx. 3 300 m ² (1)		
" water depth(1961): " 12 m		" water volume (1961): " 19 000 m ³)		
Year	Date of Formation	Date of Drainage	Kind of Drainage	Remarks
1960	In early June (2)	Probably 17 June (3)	No observation	
1961	Between 18-23 June; After first total damming up two subsequent partial lake fillings	3 emptyings: 1st: began 23 June 2nd: " 27 June 3rd: " 28 June	cf. Figs 30,31 1st: marginal (sub-marginal), later through Ice Cave 2nd: as above, but only a short time marginal 3rd: only through Ice Cave	After emptying of lake, Ermine River runs through Ice Cave during the whole summer
1962	Lake formed and emptied before observations commenced (4)			
1963	Beginning of lake filling probably a few days earlier than 1961, but com- pletion about the same time	22 June	Only marginal (sub-marginal) cutting down to the level of Ice Cave	Highest shore lines some 3 m below overflow level of 1961 (cf. Fig. 29)
1963	Lake formed and emptied before observations commenced, probably much later than 1960-1963 (5)			

(1) Estimate in Maag (1963, p. 151) too low

(2) Photo B. Robitaille

(3) Adams (1966, p. 65)

(4) Indirect evidence (personal communication C.S.L. Ommanney and H. Weiss)

(5) Indirect evidence (personal communication C.S.L. Ommanney)

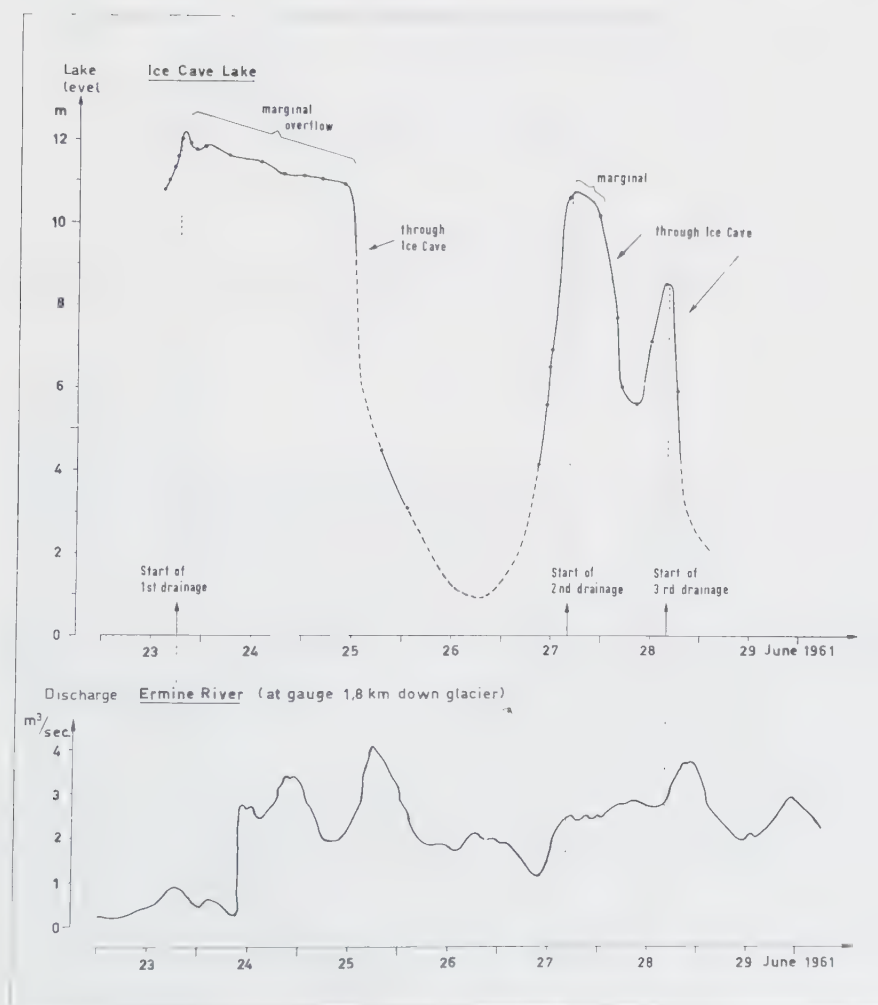


Fig. 30:
Water level of Ice
Cave Lake during
drainage and dis-
charge of Ermine
River, June 1961.

one. The channel finally reached the level of the Ice Cave and again the river ran through this old passage during the whole summer.

In 1961, contemporaneously with the observations of the drainage of the Ice Cave Lake, an automatic level recorder installed by W.P. Adams some 1.8 km down-glacier gauged the discharge of the Ermine River. It is interesting to compare the level fluctuations of the Ice Cave Lake with the discharge anomalies of the Ermine River (Fig. 30). The water volume of the Ice Cave Lake is relatively small and the outflow was rather slow, especially at the beginning when the drainage was still marginal. In addition, the meltwater channels of the previous summer had been closed during the winter at many places by the movement of the glacier, by collapsing and freezing of the tunnels or by falls of ice, snow or rock débris. Thus, a spectacular flood wave that would directly pass the gauging station a short time after the beginning of the overflow was not to be expected. But indirectly the effect of the outflow of the Ice Cave Lake was felt at the gauging station. The time lapse between the start of the first overflow of the lake and the increase of the discharge of the Ermine River registered by the level recorder was more than 15 hours. This indicates the amount of retardation the first meltwater experiences in spring and early summer proceeding along the glacier margin. The water is ponded up at many places and has first to free or by-pass the old channels and pipes and to dig new ones (Adams, 1966, p.65). Figure 31 shows how slowly the first outflow water of the Ice Cave Lake advanced down-glacier from obstacle to obstacle and from pond to pond. The sharp increase in the discharge of the Ermine River about 0930, 24 June 1961, was thus not due to the direct flood wave of the draining Ice Cave Lake but rather to the outbreak of a small marginal pond just a few hundred metres up-river from the gauging station. The outflow of this lakelet was triggered by the water from the Ice Cave Lake that finally reached this pond after several hours. The effects of the second and third filling and drainage of the Ice Cave Lake are not so obvious in the discharge curve. The water volume of the Ermine River at the gauging station during the night of 27/28 June 1961 and during the morning hours, when normally the discharge is lowest, was rather large; the level gauge also recorded some increased discharge later in the evening of 28 June. These anomalies could have been partly due to the second and third emptying of Ice Cave Lake. If this was the case, then the time lapse between the start of the outflow and the arrival of the lake water at the gauging station was much shorter than during the first emptying, which would mean that already the water had to overcome or by-pass fewer obstacles than only 4 days previously.

Phantom Lake: This is the largest ice-dammed lake in the area and one of the largest on the whole island (Table 7). Phantom Lake was formed when a tributary glacier, the Transit Glacier, retreated from the Thompson Glacier. The water is dammed up by a lobe of the latter which intrudes into the tributary valley (Fig. 32). According to Stone's stages of formation for this type of lake (Stone, 1963b, p. 343) the Phantom Lake would be at the end of the second stage, whereas the Astro Lake (Fig. 33), a similar lake in the neighbouring tributary valley, would still be in its first stage of formation.

To obtain some information on the form and depth of the still ice-covered Phantom Lake, a Kelvin & Hughes M-29 echo depth sounder (Müller, 1963i, p. 104) was used from 20 to 27 May 1960. The data have been checked at two places in 1961 and 1963 by running a sounding line through a hole in the 2 m thick ice cover at about the centre of the lake (Fig. 34). The cross-sections (Fig. 34) constructed from some 70 echo soundings and from consideration of iceberg heights show a classical U-shaped form. The irregularity in cross-section A could perhaps indicate an old moraine or a ridge separating the former Phantom and Transit glaciers which may once have coalesced here. This could only be proven with further depth soundings and bottom samplings. The longitudinal profile seems to reveal some minor steps in the lake bottom and a possible overdeepening in the western part of the basin near the damming ice lobe. Unfortunately, due to many icebergs and much drift-snow it was not possible to get more soundings in this part of the lake, and attempts to drill a hole and sound the depth with a sounding line near the ice cliff of the Thompson Glacier did not succeed. Other evidence, however, especially the height of different icebergs (Maag, 1965, p.27-28) and the thickness of overturned icebergs (Bauer, 1961, p. 18-19) speak for a possible overdeepening of this former glacier valley. In addition, the geological situation at the western end of the Phantom Lake (Fricker, 1962, Preliminary Geological Map) as well as the surface topography of the lobe of the Thompson Glacier suggest the existence of a dyke between Circle Hill and Astro Ridge, forming a kind of "riegel" at the end of the Phantom Lake valley. The Thompson Glacier valley near the junction of the tributary Phantom Lake valley is some 500 m deep (Redpath, 1965, p. 20). It seems, therefore, that the 200 m deep Phantom Lake basin is a hanging valley relative to the Thompson

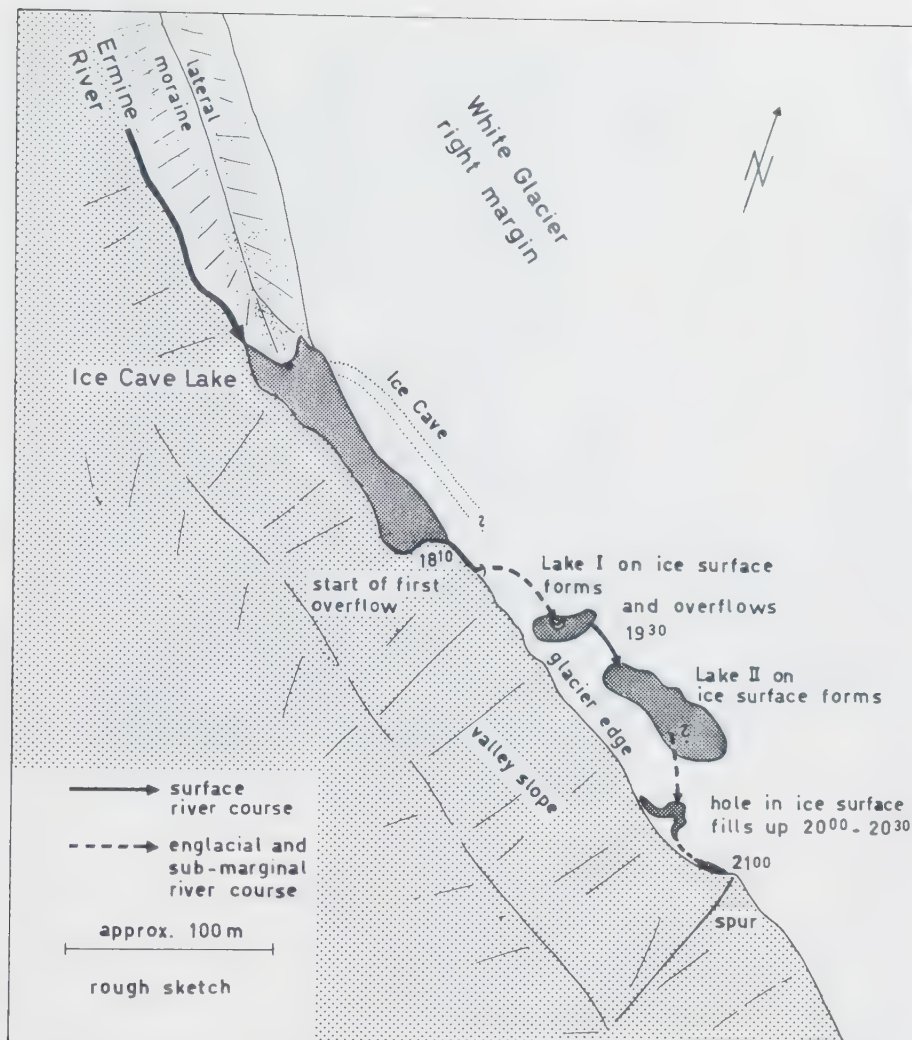


Fig. 31:
Ice Cave Lake, course
of first overflow water,
June 1961.

Table 7: Phantom Lake 1959 - 1965

Max. lake level (1960) : 397 m a.s.l.		Max. surface area (1961) : approx. $6 \times 10^6 \text{ m}^2$	
Max. water depth (1961) : approx. 200 m		Max. water volume (1961) : approx. $500 \times 10^6 \text{ m}^3$	
Year	Start of drainage	Mode of drainage	Remarks
1959	On or prior to 28 July ⁽¹⁾ ; still draining on 13 Aug. ⁽²⁾	Marginally via Raft Lake to To and Astro lakes; only small part of lake drained	After 13 Aug. and before freezing over of lake a large piece of the Thompson Glacier lobe, that protudes into Phantom Valley and dams up lake, broke off and moved as a huge table iceberg towards the centre of lake over a distance of several hundred metres (Maag, 1965, p. 9-11)
1960	7 August	As in 1959	The partial drainage was probably triggered by calving of approximately $10\,000\,000 \text{ m}^3$ of ice into lake and subsequent movement of icebergs over distances of up to 200 m (Maag, 1965, p. 11-12, 27-29)
1961	No overflow	---	Ice-cover thawed only at margin, but did not completely disappear
1962	Between 10 and 13 July ⁽³⁾	As in 1959	---
1963	No overflow	---	Completely ice-cover throughout summer, at beginning of August still almost 2 m thick (with upper 50 cm candle ice), permitting landing of a Piper Super-cub. Water temperatures in western part of the lake close to 0°C (Fig. 59)
1964	No overflow	---	Complete ice-cover
1965	Late August ⁽³⁾	As in 1959	Ice-cover remained almost the whole summer

(1) Air photographs A 16859 - 23/24; (2) Air photographs A 16864 - 34/35 (3) Personal communication C.S.L. Ommanney & F. Müller

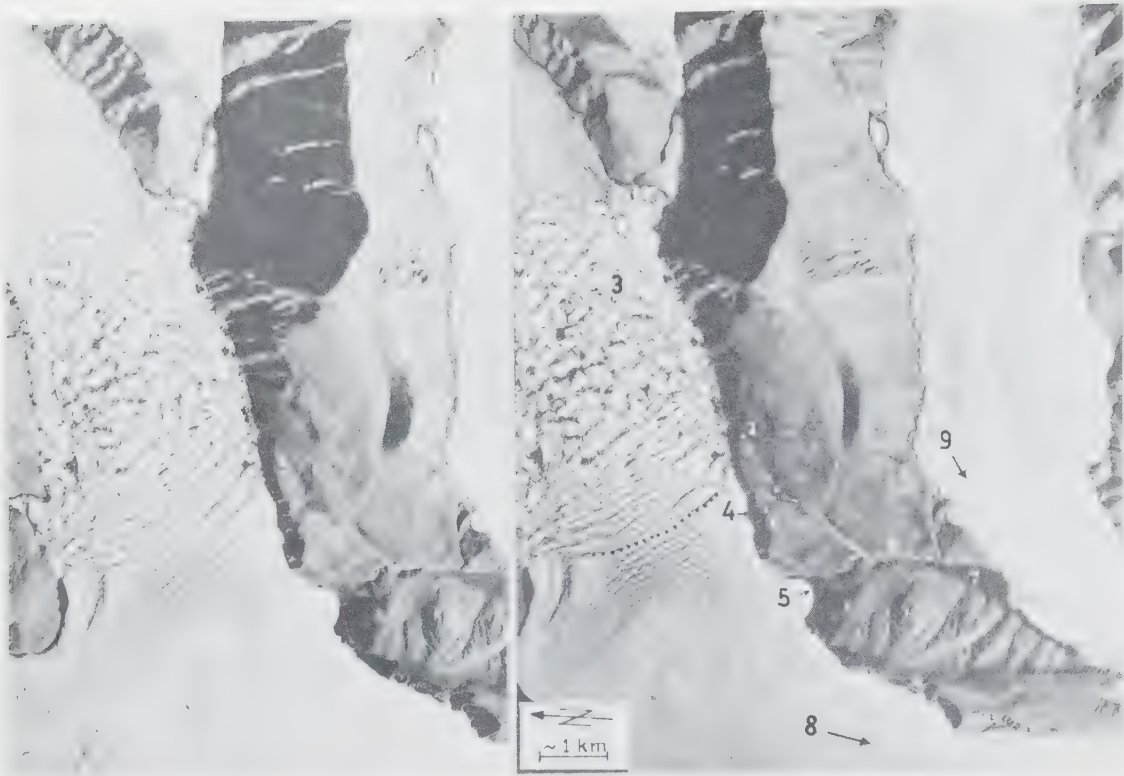


Fig. 32: Phantom Lake area. Stereopair for use with pocket stereoscope (part of vertical air photos RCAF A 16 859-24/23, 28 July 1959). 1, 2 (+ dotted white line) - old meltwater channels; 3 - Phantom Lake with icebergs; 4 - Raft Lake; 5 - To Lake; 6 - To River; 7 - N-part of Astro Lake; 8 - Thompson Glacier; 9 - Astro Glacier; 10 - Wreck Glacier. Dotted black line where damming ice lobe broke off in August 1960 (cf. Fig. 35).

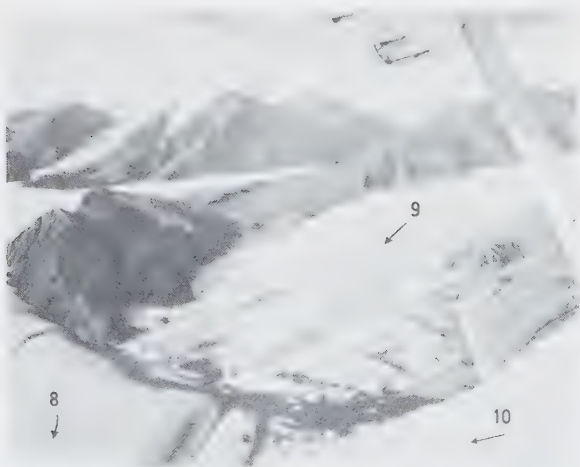
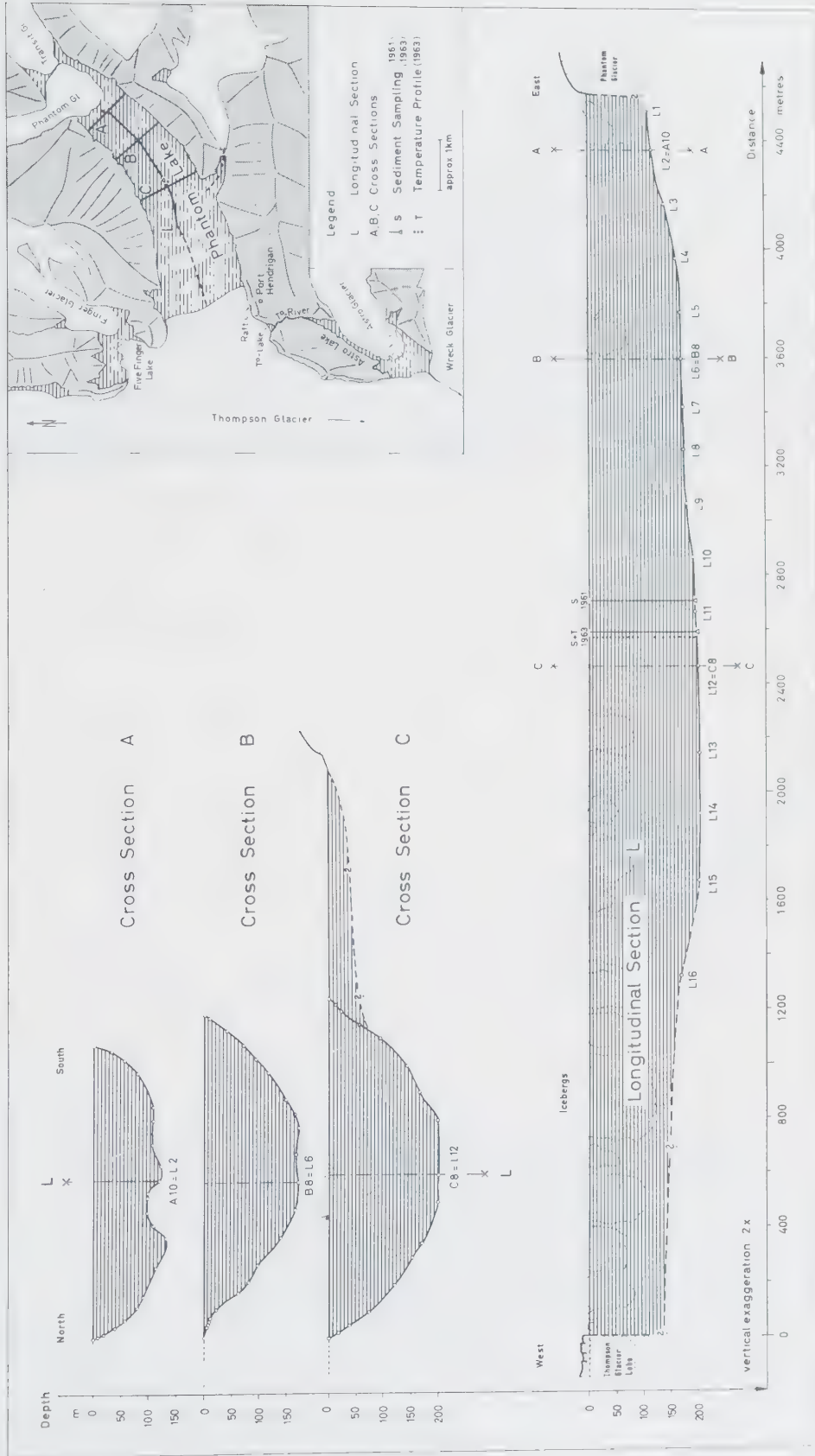


Fig. 33: Astro Lake with icebergs, dammed up between Astro (9), Thompson (8), and Wreck (10) glaciers. Phantom Lake (3) in background. (Oblique air photo, summer 1962).



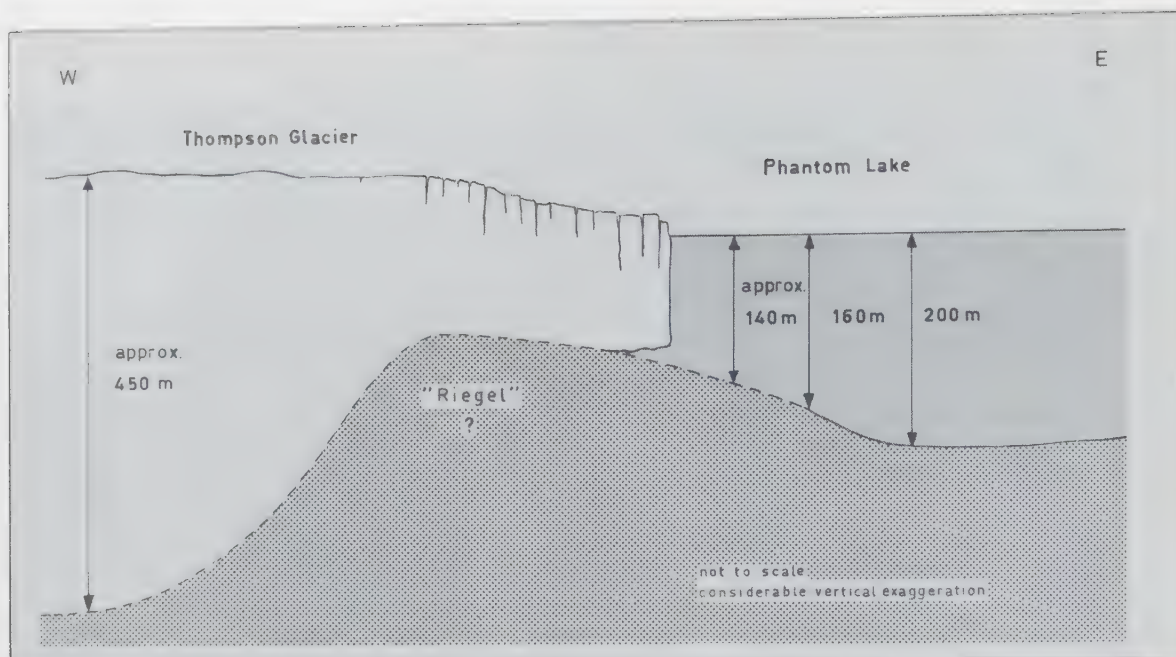


Fig. 35: Hypothetical section through Phantom Lake and Thompson Glacier.

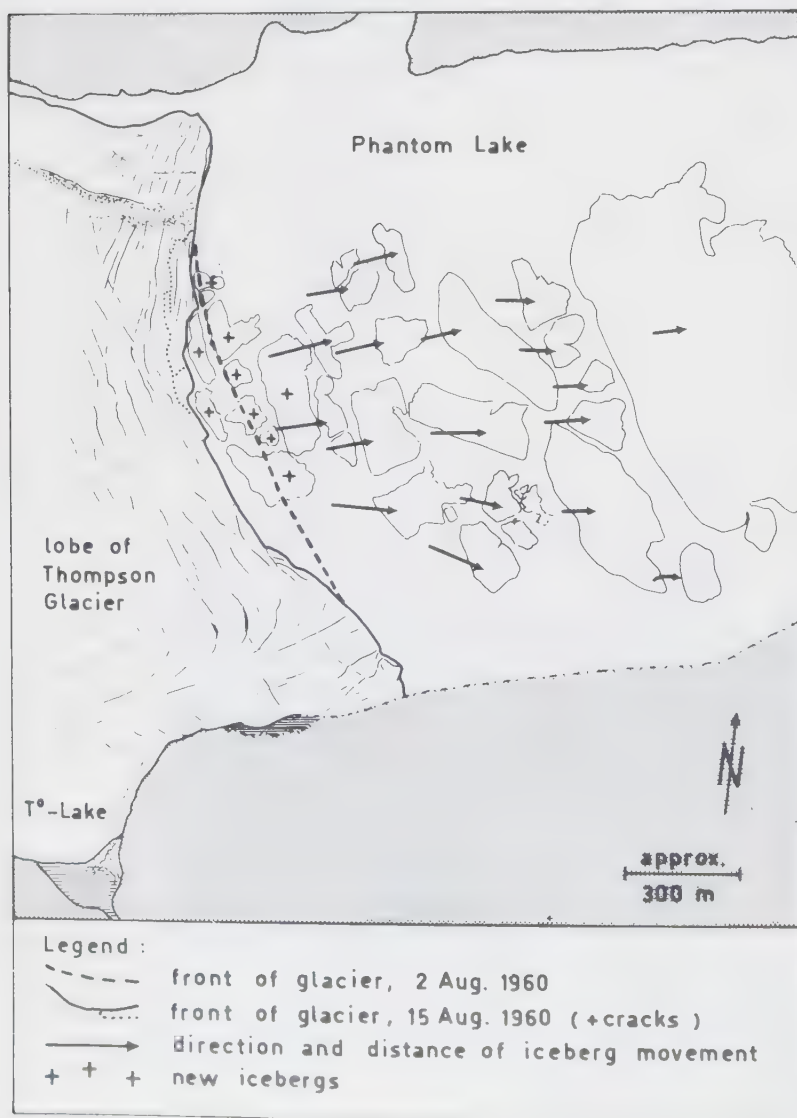


Fig. 36: Movement of icebergs after calving into Phantom Lake, August 1960. (Drawn after air photos).

Glacier trough. The hypothetical sketch (Fig. 35) attempts to show this. In Figure 34 one sees that the small tributary valley, which is now water filled and enters the Phantom Lake from the southeast near cross-section C, is probably also a hanging valley but relative to the Phantom Lake basin. It is a good example for a structurally controlled valley and lake basin located along a faultline (Fricker, 1962, Preliminary Geological Map).

Phantom Lake is the most prominent lake in a whole system of interdependent ice-dammed lakes (Figs 32, 34): lakes dammed up by the Parallel Glacier --> lakes at the foot of Ore Mountain and Tooth Ridge --> Five Finger Lake --> Phantom Lake --> Raft Lake --> To Lake --> Astro Lake. If the connections between these lakes function - which is the case every year - the overflow of a higher lake may trigger the drainage of lower ones. Table 7 gives data concerning the drainage of Phantom Lake. It does not drain every summer, and when it drains only a small part of the water, some 1 - 2%, flows out and the lake remains practically full over winter, freezing completely over. During the period 1959-65, the earliest overflow was observed in 1962 when Phantom Lake started to drain some 1 to 2 weeks earlier than in 1959 and almost 4 weeks earlier than in 1960. In 1961, 1963 and 1964 it did not overflow at all. In 1959, 1960 and 1962 the lake drained in the same way, namely by overflowing marginally to Raft Lake and from there to To Lake and Astro Lake. The existence of a large marginal channel between Raft Lake and To Lake and a wide overflow channel between To Lake and Astro Lake, visible on the aerial photographs of 1959, suggests that already prior to 1959 Phantom Lake overflowed in a similar way and helped to erode these river beds. If there was besides this marginal drainage also some sub- or englacial outflow, and why only a small percentage of the lake water flowed out is discussed in the next chapter.

Direct observation and the study of aerial photos suggests that the breaking off of huge amounts of ice into the lake, pushing the already present icebergs over large distances (Fig. 36) and producing high waves, triggered the overflow of Phantom Lake in 1960. In the previous summer, on the other hand, substantial calving must have taken place after the start of the lake drainage (Maag, 1965, p. 6-14).

c) Discussion of Drainage Mechanism

"Normal" drainage: The field observations as well as the study of aerial photographs show that most of the lakes in the Expedition Area drain by flowing over the damming ice (like Between Lake), by spilling over between glacier margin and valley slope (like Crusoe-Baby Lake) or across a threshold situated lower than the damming ice. An outflow channel is cut by melting, and mechanically, from the surface down into the glacier ice, or between valley slope and ice, or into the ground. Overflowing the ice barrier and melting down an outlet channel has been called "the simplest way imaginable for the drainage to start" (Listøl, 1956, p. 123). That the melting activity of outflowing lake water can be very important in the drainage process of ice-dammed lakes has been recognized by various scientists, although mostly for lakes in regions with temperate glaciers (de Saussure, 1804, p. 397; Finsterwalder, 1897, p. 6-11, in Heinsheimer, 1954a, p. 29; Lütshg, 1915, p. 205-206; 1950, p. 34; Stone, 1955, p. 74; Liestøl, 1956, p. 123-125, 130; Charlesworth, 1957, p. 453; Lanser, 1959, p. 28-29; Marcus, 1960, p. 103-106).

The following rough calculation, which is similar to the one Liestøl (1956, p. 124-125) used, gives an idea of the melting capacity of the supraglacially outflowing Between Lake water: the volume of water flowing out of Between Lake, until the lake is 95% empty, amounts to about $10\,000\,000\text{ m}^3$ (some $7\,000\,000\text{ m}^3$ stored in the lake basin + some $3\,000\,000\text{ m}^3$ flowing into the basin during the emptying). In 1963, the outflowing water had an average temperature of about 2°C . To melt 1 gram of ice (at 0°C) requires 80 cal and for 1 gram of cold ice (e.g. at -10°) some 85 cal (assuming a specific heat of about $0.5\text{ cal g}^{-1}\text{ }^\circ\text{C}^{-1}$). Thus, $1\text{ m}^3 (=1\,000\text{ kg})$ of water of 2°C could theoretically melt some 23 kg of ice. The conversion of potential via kinetic energy into thermal form may be another important source for melting ice. The height difference from the Between Lake surface down to the end of the glacier snout is about 100 m (Fig. 37). When 1 m^3 of water is running down from Between Lake to the snout some 230 kcal of heat could develop. Theoretically, almost 3 kg of ice could be melted with this heat, which would increase the melting capacity of 1 m^3 of Between Lake water to about 26 kg of ice. As this is only an estimate, the amount of heat lost or not used, as well as the problems of heat transfer from water to ice, of heat conduction in cold ice, of heat gain due to the inner friction of turbulent water and due to condensation, etc., are not discussed here. But even if we assume that only some 50% of the available heat is applied for melting, the outflowing $10\,000\,000\text{ m}^3$ of water could melt some $130\,000\text{ m}^3$ of ice, or a channel

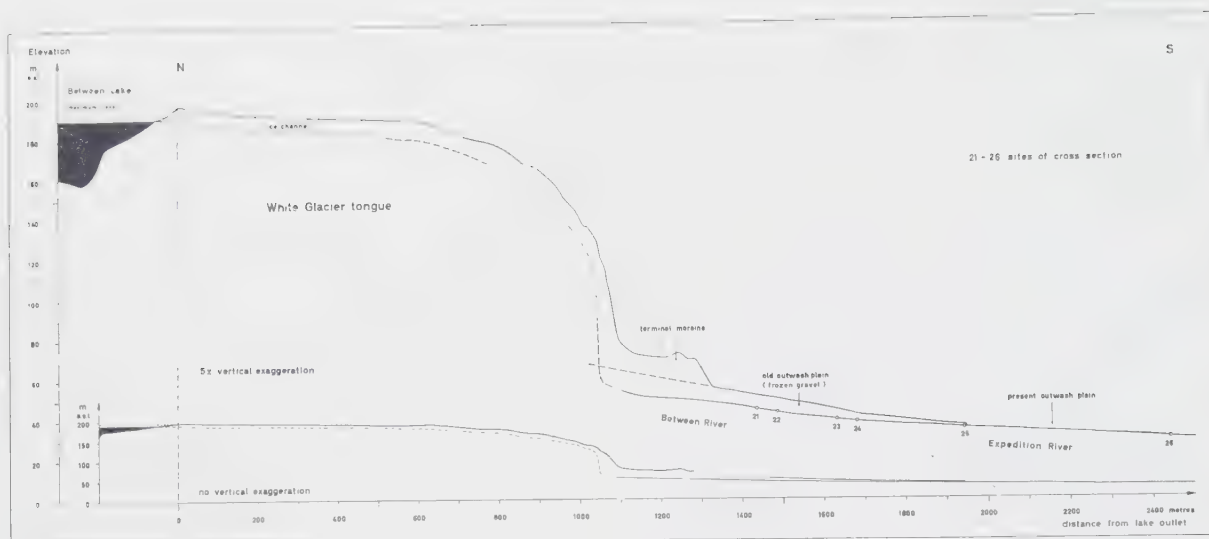


Fig. 37: Longitudinal section Between Lake - Between River.

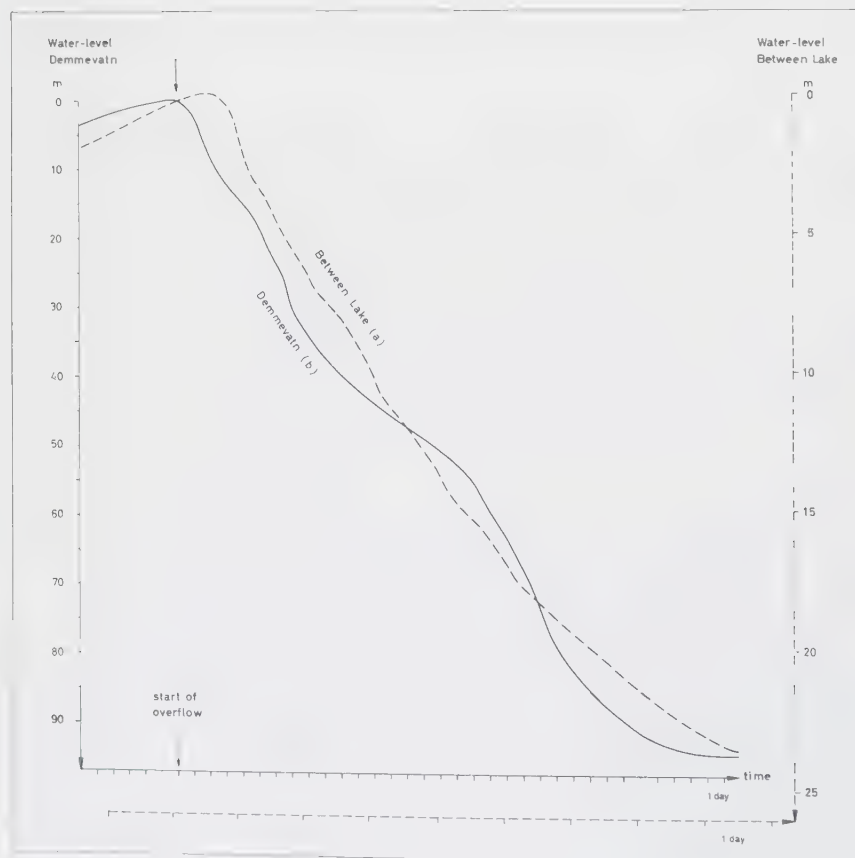


Fig. 38:
Emptying curves of two
supraglacially draining
lakes: a) Between Lake
(Axel Heiberg Island),
July-August 1961, and
b) Demmevatn (Norway),
August-September 1897.
(After Liestøl, 1956,
p. 129, Fig. 4)

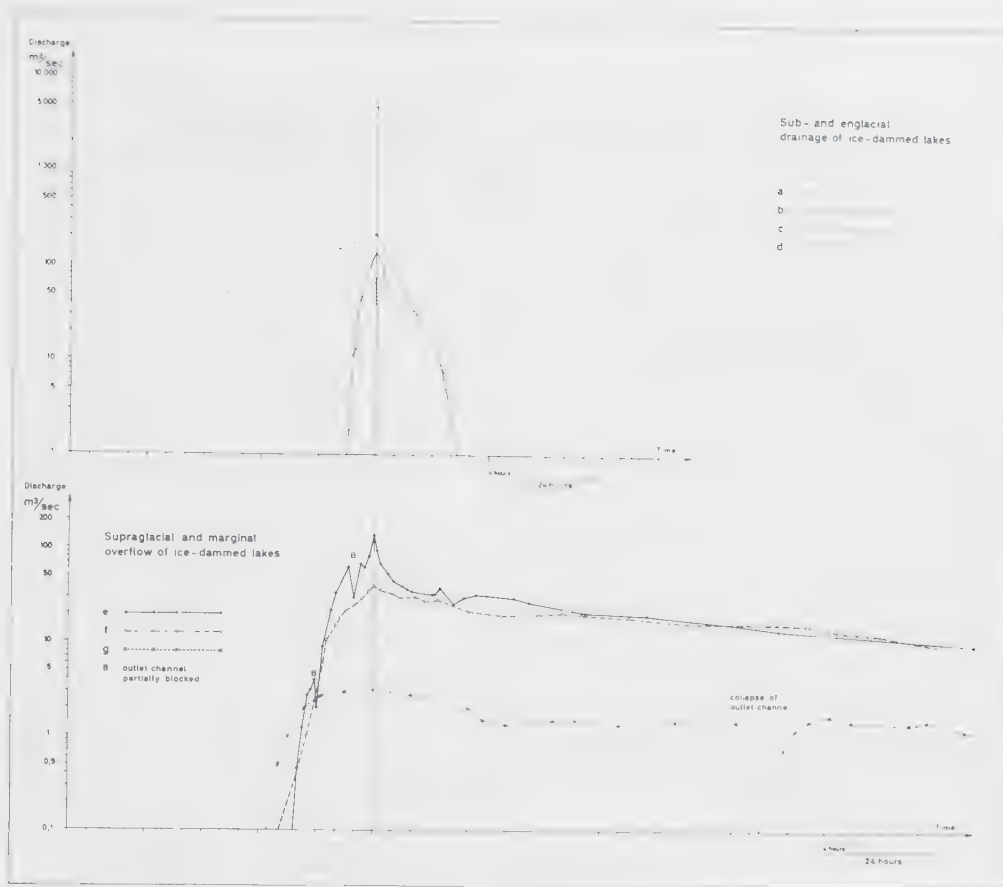


Fig. 39:

Discharge curves of different ice-dammed lakes: a) Graenalon, and b) Gjanupsvatn (Iceland), after Thorarinsson, 1957, p.23; c) Marjelen See (Swiss Alps), after Lüttschg, 1915, p.272,296; d) lake dammed by Mitdluagkat Glacier (SW Greenland), after Larsen, 1959, plate I; and e) Between Lake 1963, f) Between Lake 1961, g) Crusoe-Baby Lake 1961 (Axel Heiberg Island).



Fig. 40: Waterfall emerging from Thompson Glacier snout, 10 August 1963, 1300; man in circle for scale.

about 1000 m long, 5 m wide and 25 m deep. This estimate agrees well with the observations of 1962 when a whole new ice channel of similar dimensions was cut by the outflowing Between Lake water during one emptying. From the above it is apparent that the lake water temperatures must not be much above 0°C to guarantee the rapid deepening of the outlet channel. It is evident, too, that during the 1963 drainage of Between Lake the "surplus heat" was relatively much more important than the heat gained from the conversion of potential energy; the ratio was about 7:1.

The great contribution of melting to the enlarging of the outlet channel of Between Lake and the minor importance of other forms of fluvial erosion is evident especially in the first, more gently sloping section. Here, the outflowing water still has - even a few hundred metres downriver from the outlet - a large enough "heat surplus" to effectively melt the ice. The result is a more or less straight channel with rather smooth walls (Fig. 19). On the other hand, in the last section of the channel cut into the White Glacier tongue the "heat surplus" of the draining water is just about exhausted. The channel is less straight and displays erosional forms which suggest that in this much steeper section erosion by the highly turbulent fast flowing water plays a more important rôle.

Variations of the lake water temperature may be one of the main reasons for differences in the speed and completeness of the drainage from year to year and from lake to lake. It has already been mentioned that the relatively low temperature of the Crusoe-Baby Lake water might have been one of the reasons for the rather slow outflow. Similarly, the variation of the speed of the different Between Lake emptyings might in part be due to yearly differences in the average lake water temperature. Again, the situation at the outlet of Phantom Lake may become clearer from consideration of the above calculations. In August 1963, the water of Phantom Lake approached 0°C (4 of this Chapter); that means that the melting capacity of this cold water was very limited, and this was not much different in the summers when Phantom Lake overflowed. In addition, the heat supply gained from the conversion of potential energy is very small because of the minimal height difference between Phantom Lake and Raft Lake. This lack of heat for melting may be the main reason why only a small part of the lake emptied and why the drainage stopped after a certain period of overflow. It seems that the enlarging of the outlet channel was relatively slow, probably slower than the closing of the channel by the sideward push of the Thompson Glacier lobe. This explanation would be in agreement with Liestøl's statement (1956, p. 123) that "the heat surplus (in arctic lakes) will often be sufficient for the drainage only of some few metres during summer". And yet, it must be stressed that the temperature régime of Phantom Lake is more or less exceptional in the Expedition Area and representative - if at all - only for lakes with an abundance of icebergs, whereas the situation in Between Lake and Crusoe-Baby Lake, described above, is the more common.

It is interesting to compare the speed and progress of the drainage of the lakes investigated in the Expedition Area with similar observations elsewhere. Several authors state that supraglacial and marginal drainage of ice-dammed lakes proceeds much more slowly than sub- and englacial drainage, and that there is often just a partial depletion (Finsterwalder, 1897, p. 6-11, in Heinsheimer, 1954a, p. 29; Lütschg, 1915, p. 204; Collet, 1925, p. 28; Arnborg, 1955b, p. 220; Liestøl, 1956, p. 123, 128-130; Weidick, 1963, p. 112). The observations at Ice Cave Lake (Chapter B2b) seem to corroborate this. In that case the water did not first have to erode a new tunnel but could utilize the old one, once it had removed the plug. Liestøl (1956, p. 123) states that with the melting down of the outlet channel through the ice barrier "the depletion of the lake gradually takes place". A more or less gradual emptying was observed also in the Expedition Area, although with a rapid increase of the discharge during the first phase. However, as discussed before, the speed of the depletion can differ from lake to lake and from emptying to emptying, largely due to differences in the lake water temperature. Lakes overflowing marginally might drain a little more slowly and perhaps less evenly than those emptying over the glacier surface, mainly because of the more frequent collapses of the (sub-) marginal outlet channels. Despite these variations in speed the general form of the emptying curves of all the lakes investigated is roughly the same. There is also an interesting similarity between the emptying curves of Between Lake and Demmevatn, a Norwegian ice-dammed lake (Liestøl, 1956, p. 129), both of which drained supraglacially (Fig. 38). In Figure 39 an attempt is made - despite the great variations in size and form of the lake basins - to compare the discharge curves of Between Lake and Crusoe-Baby Lake with some curves of ice-dammed lakes, dammed by temperate glaciers and emptying subglacially. Thorarinnsson (1939b, p. 227) describes the development of the discharge during a subglacial emptying, a 'jökulhlaup' of Graenalón, a large "limno-glacial" ice-dammed lake in Iceland, as "slowly at first and then at an ever increasing rate until the run-off culminates, after which the spate diminishes very rapidly".

Also Arnborg (1955a, p. 196) states that "the lowering from maximum to normal value after the 'jokulhlaup' takes place about 4 times as rapidly as the rising from normal value before the maximum". Larsen (1959, p. 63) observed a similar type of discharge in southern Greenland; he describes it as a "relatively gentle increase until the maximum value, and then a sudden drop". It is evident from Figure 39 that after a rapid increase of the outflow volume up to maximum discharge, instead of a sudden decrease as reported from various subglacially draining ice-dammed lakes, the supraglacially draining Between and Crusoe-Baby lakes showed a more gradual decrease. Thus, it appears that subglacial and supraglacial depletion of ice-dammed lakes progresses differently.

Many lakes drain annually (e.g. Crusoe-Baby Lake), others only in summers with a good meltwater supply (e.g. Phantom Lake). There is no evidence in the Expedition Area that the small lakes empty less frequently than the large ones, as was observed by Stone in Alaska (1963b, p. 338). On the contrary, the largest lakes in the area, like Phantom Lake, Five Finger Lake, Astro Lake, usually empty only partially and that not every year. But this is probably more dependent on the water temperature than on the size. The drainage frequency does not depend on the absolute size of a particular lake but on its relative one, i.e. on the relation between the water supply, which can vary considerably from year to year, and the volume of the lake basin. Most emptyings seem to take place during July and early August, but the date depends very much on the weather of that particular summer (Chapter D). So far, no drainage of an ice-dammed lake in the Expedition Area during winter has been recorded; this is rare also in temperate glacier regions (Lütschg, 1950, p. 34; Stone, 1955, p. 66; Liestøl, 1956, p. 122; Marcus, 1960, p. 95, 98).

A lake ponded by cold ice starts to drain only when it is full. But the actual overflow can be triggered and accelerated in various ways, i.e., by rain, which is usually rather warm and runs off immediately (as it cannot penetrate into the cold glacier or the frozen ground) so that an almost full lake basin receives in a short time enough additional water and heat to be able to overflow (e.g., the triggering of the outflow of Crusoe-Baby Lake in 1961); by waves produced by the breaking off of icebergs (Fig. 36) or the sliding of large amounts of debris from the valley slopes into the lakes, or high winds (e.g., the start of the drainage of Phantom Lake in 1960); or by the outflow of lakes situated up-glacier (e.g., the outflow of a marginal lake on the upper Thompson Glacier and of Five Finger Lake bringing the 1960 Phantom Lake to an overflow which in turn triggered the outflow of Raft, To and Astro lakes). On the other hand, the overflow might be delayed as, for instance, by the temporary piling up of pieces of lake ice or icebergs (e.g., Between Lake in 1965).

"Abnormal" drainage: As well as the "normal" supraglacial and marginal emptying of ice-dammed lakes some instances of possible en- or subglacial drainage have been observed, or presumed, in the Expedition Area.

At Ice Cave Lake the water does not use a true subglacial drainage route, but rather a buried marginal passage which is usually freed by the hydrostatic pressure and the melting activity of the spring meltwater ponded up at this site. As the margin of the White Glacier near the Ice Cave is relatively quiet and rapidly thinning (Müller, 1963g, p. 71-74), probably the near-marginal tunnel as well as some englacial pipes further down-glacier can remain open more easily than in a more dynamic part of the glacier.

In the summers of 1960 - 1963 it was observed that the largest influent from the White Glacier entered the Between Lake through an englacial pipe; both position and water volume varied somewhat from year to year. The water usually gushed out with some pressure showing some hydrostatic overhead. The exact intake of this pipe has not been located so far as the planned dye-tracing had to be cancelled, but it probably lies only a few hundred metres up-glacier. In 1964, C.S.L. Ommanney observed water coming out of the White Glacier snout from what appeared to be a relatively large englacial stream. In 1961 and 1963 outflows of englacial water had been observed near the terminal moraine of the White Glacier and also at the junction of the snouts of the White and Thompson glaciers. The gradient of the englacial temperatures measured at the Lower Ice Station in the tongue area of the White Glacier (Müller, 1963h, p. 87) suggests that at the bottom of the glacier the ice temperature is at 0°C; the hypothesis that meltwater is present at the base of the lower White Glacier has been forwarded by Müller (1963g, p. 71). Towards the end of the melt season (13 August 1963) after the Between Lake had completely drained, the water collected at the ice front where the Between River leaves the White Glacier contained relatively large amounts of dissolved substances, especially a surprisingly high concentration of sulphates, in comparison with the water collected earlier in the summer (Table 9). Fricker (1963a, p. 142) suggests

that below the White Glacier tongue there may be some evaporite, the sulphate concentration could then result from water flowing subglacially in contact with the gypsum. The possibility, therefore, of part of the Between Lake water occasionally finding sub- or englacial passages had to be considered. However, a close inspection of the filling and emptying curves and a comparison with weather and ablation data and with discharge values did not reveal any hidden sub- or englacial leaking. Furthermore, no englacial stream intakes were found in the empty lake basin but only the supraglacial outlet channel. The lake remained filled during the winter 1964/65 without any leaking. Moreover, the whole lake bottom is underlain by impermeable permafrost. Even the deepest parts of Between Lake lie only some 40 m below the surface of the damming glaciers whereas the White Glacier is still more than 100 m thick near the Lower Ice Station, which means that the ice layers ponding the Between Lake water are clearly of negative temperature. All these facts suggest that - at least in the period from 1959 to 1965 - Between Lake drained only supraglacially and did not lose any water through sub- or englacial passages.

At several places on the Thompson Glacier, in the middle part of the ablation area and also in the tongue area, supraglacial rivers disappearing into moulin-like holes were observed. Some of these moulins are indicated in the topographical maps of 1:50 000 (Thompson Glacier Region, 1962) and 1:5 000 (Thompson Glacier Snout, 1962). In 1963, 1964 and also in 1965 a strong waterfall came out of the upper part of the Thompson Glacier cliff although not always at the same place (Fig. 40). Previously in 1960 and 1961, waterfalls, but much smaller ones, had been observed. The position and the volume of these waterfalls changed from year to year or even during one summer. In the ice wall of the Thompson Glacier snout there are several abandoned water pipes now filled with sand. No correlation between the discharge of the waterfalls and the emptying of ice-dammed lakes further up-glacier has been detected so far. Future investigations should try to trace these presumably englacial water courses of the White and Thompson glaciers. None of them seem to be very long. They might follow old supraglacial channels or crevasse systems which were closed only on the surface. Some of them might run along internal moraines or the borders of inset or superimposed (Sharp, 1947, p. 28) glacier sections.

During the partial outflow of Phantom Lake in August 1960 it was observed that besides the water that flowed marginally from Phantom Lake to Raft Lake (Fig. 34) there seemed to be some, but not much, water entering the Raft Lake subaqueously through an opening in the lake basin. The Raft Lake basin is situated mainly on the marginal ice of the Thompson Glacier lobe. Early in 1961, before this lake basin was again filled with water, a narrow, crevasse-like opening of some 4 m length was seen at the site where in 1960 some water probably entered through the lake bottom. Very likely this water was just meltwater from the surface of the lobe that disappeared near the margin into a crevasse and found its way to Raft Lake by a short englacial pipe. However, the possibility of sub- or englacial drainage of Phantom Lake as well as the marginal overflow has to be considered. The mechanics of sub- and englacial emptying of ice-dammed lakes is still the subject of much discussion and speculation, especially because of the difficulties of direct investigations (Liestøl, 1956, p. 122; Marcus, 1960, p. 98-99; Stone, 1963b, p. 341; Mathews, 1965, p. 52). Good summaries and discussions of the various theories have been given by Nichols and Miller (1952, p. 43), Glen (1954), Liestøl (1956, p. 123-125), Charlesworth (1957, p. 453, 458), Marcus (1960, p. 99-104), and Stone (1963b, p. 340-341). It would appear that none of the proposed hypotheses exactly fits the case of Phantom Lake. If sub- and englacial drainage of Phantom Lake does occur it is probably due to a combination of different processes, e.g. of seepage (Carey and Ahmad, 1961, p. 882; Andrews, 1964, p. 119) supplemented by the enlarging of tunnels as proposed by Glen (1954) and by melting (Liestøl, 1956, p. 123-125) and perhaps also assisted, or simply triggered, by hydrostatic lifting (Thorarinsson, 1939b, 221-222) in the modified sense as cited by Marcus (1960, p. 102-103). But, considering that 1) the Thompson Glacier lobe damming up Phantom Lake is a cold glacier, 2) the crevasse pattern and the observations suggest only very limited and temporary floating of small parts of the glacier lobe, 3) the temperature of the Phantom Lake water is very near 0°C (Chapter B), therefore, its melting capacity is very small, and 4) there exists possibly an obstacle, a "riegel", between the Phantom Lake valley and the Thompson Glacier valley (Chapter B, Fig. 35), it is very unlikely that sub- or englacial water reaches the Thompson Glacier valley from the lake, even if we allow the possibility of some water penetrating a short distance under the damming ice lobe, thus perhaps bringing the temperature of this part of the ice and of the ground to near the pressure melting point.

It should be stressed, however, that in spite of this relatively lengthy consideration of possible en- or subglacial drainage the ice-dammed lakes in the Expedition Area empty normally

by overtopping of rock or ice barriers. This differs from observations on Alaskan glaciers where subglacial lake drainage along marginal, non-marginal and, possibly, englacial openings is the rule but where there is no evidence of emptyings by overflowing the damming ice (Stone, 1963b, p. 340-341). The glaciers of Stone's investigation area are of Ahlmann's temperate glacier type whereas those of the Expedition Area are sub-polar. This might explain the differences in the mode of lake emptyings. The cold ice of the Axel Heiberg glaciers and the permafrost of the valley slopes do not normally allow the lake water to drain sub- or englacially, or through the soil. This is in agreement with observations of Schytt (1956, p. 74) on the emptyings of two lakes dammed by sub-polar type glaciers near Thule, northwestern Greenland, drained normally by supraglacial streams. There are various other reports of predominately supraglacial and marginal or sub-marginal drainage on glaciers with cold ice in northern Greenland, Baffin Island, Ellesmere Island and Axel Heiberg Island (Ward, 1952, p. 9-11; Nobles, 1954, p. 1290; 1961, p. 760-761; Smith, 1961, p. 230; Falconer, 1962, p. 4, 26; Robitaille and Greffard, 1962, p. 86-87; Maag, 1963, p. 155; Rudberg, 1963a, p. 214; 1965, p. 4; Ives and Kirby, 1964, p. 920). Thus Stone's statement that probably most of the world's ice-dammed lakes empty through sub- or englacial channels (Stone, 1963b, p. 340) may be true for temperate glaciers but not for those of sub-polar type. There are some reports even from temperate glaciers of marginal and supraglacial drainage of ice-dammed lakes but these are exceptions to the normal sub- or englacial process (e.g. Lüttsch, 1915, p. 178-182, 206; Heybrock, 1934, in Stone, 1963b, p. 340; Liestøl, 1956, p. 123, 128-130; Charlesworth, 1957, p. 458-459). It thus appears that, besides the dynamics of the glacier and some other factors (Mannerfelt, 1945, p. 15-16; Marcus, 1960, p. 106; Stone, 1963b, p. 341), it is to a high degree the geophysical nature of the ice (temperate or cold) that dictates the type of drainage of ice-dammed lakes.

d) Some Geomorphological Consequences of Lake Level Fluctuations

Some minor ancillary observations on the effect of the filling and emptying of ice-dammed lakes were made. They corroborate findings reported from other glacierized or once glacierized regions.

Shoreline features: Flint (1957, p. 145) writes that the short duration of glacial lakes together with fluctuations of the water level might be reasons for the non-existence of shore features. It is obvious that shore features cannot develop as well with periodically draining ephemeral ice-dammed lakes as with enduring lakes. An illustration of this is Ice Cave Lake, which fills for just a few days every year, where the shorelines - especially on the ice - are only faintly visible after drainage and disappear in a short time. However, at Between Lake distinct shore features are formed (Figs 13, 16), and it was noted that the shorelines visible in the aerial photographs of 1959 were still present in 1963 although the lake filled and emptied 3 times in the interim. Also, for many ice-dammed lakes of the Expedition Area old elevated shorelines and watermarks could be seen from the air or in aerial photographs; some of them might be just "lichen-trim zones" (Ives, 1962, p. 200; Ives and Andrews, 1963, p. 21) which can also reflect former water levels (Beschel, 1961b, p. 1049).

The distinct shoreline features observed at Between Lake and other ice-dammed lakes can be formed in various ways:-

(1) The lakes in the Expedition Area do not generally empty sub- or englacially but fill up until they can overtop the barrier. This means that the maximum lake level does not change much from year to year; exceptions are lakes dammed up by rapidly thinning and receding glaciers. Due to the stability of the maximum lake level over a number of years a relatively well developed shoreline can form. This maximum shoreline is not necessarily the best developed one as the lake usually remains at this maximum level only a relatively short time. Equally, well developed fossil shorelines do not always represent former maximum lake levels.

(2) Marked shore features, lower than the present highest ones, may be formed during stagnation of the lake level following a collapse of the outlet channel; such were observed at Between Lake and Crusoe-Baby Lake.

(3) Some of the Between Lake and Crusoe-Baby Lake shorelines may be due to excessive wave action during stormy periods.

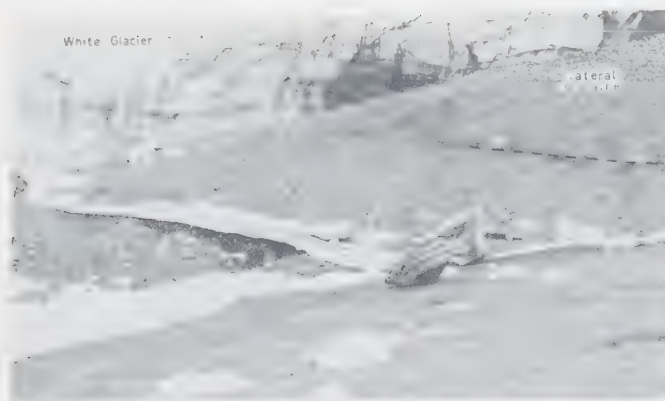


Fig. 41: Remnants of ice that covered Between Lake in winter 1964/65; ice rim left after lake drained in summer 1965 - dotted line 10 m above indicates maximum lake level 1965 (29 July 1965).

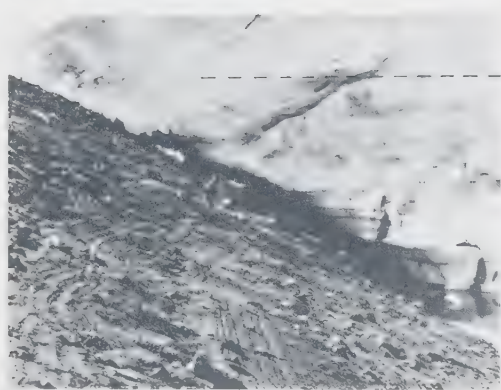


Fig. 42: Fine shorelines in basin of Crusoe-Baby Lake, 22 July 1961. Upper part of outlet channel between land and glacier margin is collapsed. Dotted line indicates highest water level 1961 (5 July).

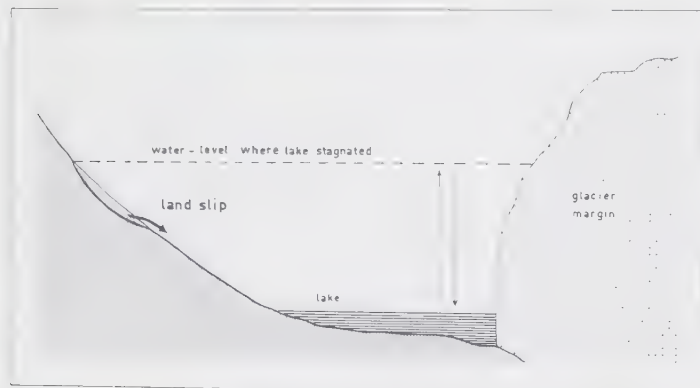


Fig. 43: Diagrammatic cross-section through basin of ice-dammed lake with landslide.



Fig. 44: Dissected delta in basin of Crusoe-Baby Lake (22 July 1961); sub-marginal outlet channel in background. Dotted line shows highest water level 1961 (5 July) some 14 m above present level.

(4) The lake ice cover that formed during the winter 1964/65, when Between Lake remained full, left a rim of ice still frozen to the shore after the lake had drained in the summer of 1965 (Fig. 41). Similar ice-rims have been observed at Raft Lake and Phantom Lake in 1961. The melting of such an ice-rim together with the effect of the pushing lake ice could produce a marked line on the slope. It may be that some of the small beaches at Between Lake, situated lower than the maximum shoreline, result from the melting of former such ice-rims.

Whether or not the less well defined shorelines are preserved depends mainly on the kind of shore material. In the relatively coarse material of the higher parts of the Between Lake basin closely spaced fine shorelines are not usually found, whereas on the shale slopes of the Crusoe-Baby Lake basin whole series of them were observed (Fig. 42). These micro-shorelines appeared to result from a steplike lowering of Crusoe-Baby Lake. A comparison of the spacing of these shorelines with the speed of the lowering of the water level suggests that the majority of them reflect a daily rhythm. The amount of inflowing water varies usually between day and night, whereas the volume of the outflowing water was observed to remain almost constant. Therefore the water level falls slowly or stays unchanged during afternoon and evening when the maximum inflow occurs, and falls faster during the latter part of the night and the morning when the inflow volume is smaller. Thus a fine shoreline forms.

Shorelines cut into the face of the damming glaciers by the thermal and mechanical action of the relatively warm surface water (Lütschg, 1915, p. 25; de Boer, 1949, p. 336; Heinsheimer, 1954b, p. 37; Stone, 1955, p. 52) were also common features in the Expedition Area (Figs 13, 24). These usually disappeared after a few days but they were useful in the reconstruction of certain lake levels despite their very ephemeral character.

Effects of icebergs and lake ice: The disturbance of lake deposits and shorelines by the drag and push of grounded icebergs and lake ice pieces and their melting, as described by Okko (1956, p. 56-57), Flint (1957, p. 145), Røen (1962, p. 63-65), Fulton (1965, p. 564, 566) and others has also been observed in the basin of Between Lake, Crusoe-Baby Lake, and others (Figs 13, 44). The effect of the broken-up lake ice cover and the many icebergs that were stranded in the Between Lake basin in the summer of 1965 must have been very great (Fig. 41).

Disturbance of slopes by submergence: The saturation of the surface layers during the submergence of the steeper slopes of a lake basin may speed up solifluction and sub-surface flow and may be the main reason for the occurrence of small landslips, slumping and mudflows which were observed to start frequently from the high water level or from levels where the lake had stagnated for a longer period (Fig. 43). All these mass-wasting processes disturb or even destroy shore features and stratified lake deposits (Flint, 1957, p. 145). Whether or not the removal of the stabilizing forces exerted by the water masses of a lake (Ward, 1953, p. 247) may cause slides or rock and ice-falls has not yet been studied in the investigation area.

Change of mouths of inflowing rivers: Dissected deltas and trenched sediments in ice-dammed lakes have been mentioned by various authors (Arnborg, 1955b, p. 220). Recent examples of such features as well as elevated former ones are quite numerous in the Expedition Area (Fig. 44). Their formation is complicated because in periodically draining lakes rapid changes of the local base level of erosion occur within a short time. This gives rise, at one particular place, to a complex sequence of accumulation and erosion periods in the course of one summer and over the years. The complexity is increased by the fact that, besides the change of the base level, the inflow water volume and the sediment load fluctuate considerably during each day and throughout the summer. The constant changes in a lake with consecutive fillings and emptyings are also reflected in the recent modifications of the river course of the main influent from the White Glacier to Between Lake (Figs 45-47).

Deflation of dry lake sediments: Lake deposits, especially of small grain size, mentioned by Wegmann (1935, p. 54) and Stone (1955, p. 10) as being blown out of empty lake basins were observed at Between Lake, Crusoe-Baby Lake and some small lakes along the margins of the White and Thompson glaciers. The areas most affected were the high and bare ones which were dry earlier and longer than the lake bottoms. Deflation is not equally effective every year, e.g. in 1963 only a very few parts of Between Lake dried out sufficiently to permit this wind action. Such deflation, along with the washing out of the finer deposits by wave and current action and by the outflowing water, is largely responsible for the relative scarcity of the smaller grain sizes in comparison with the sediments of lakes which do not drain periodically (see 3a and 3 b below).

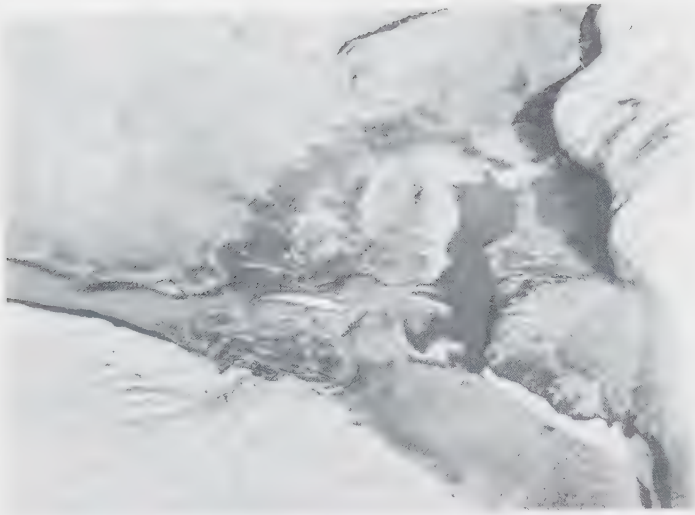


Fig. 45: Between Lake basin, summer 1962, with various shorelines. Influent river in foreground cutting through lateral moraine of White Glacier (oblique air photo).

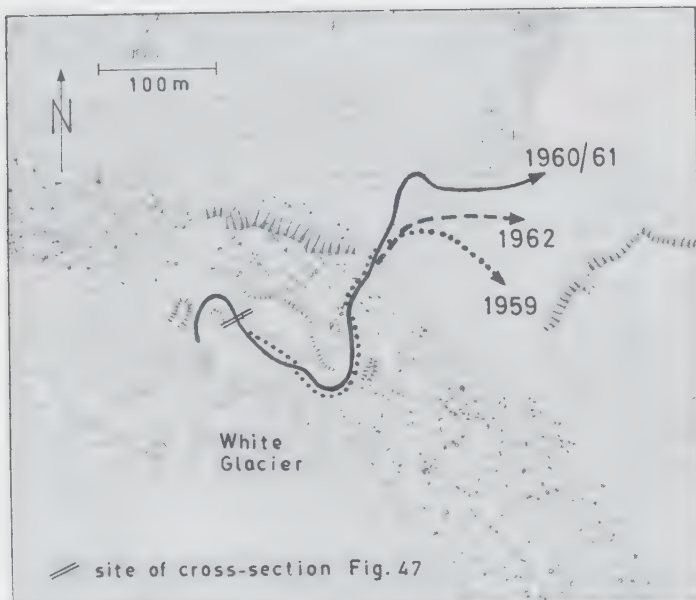
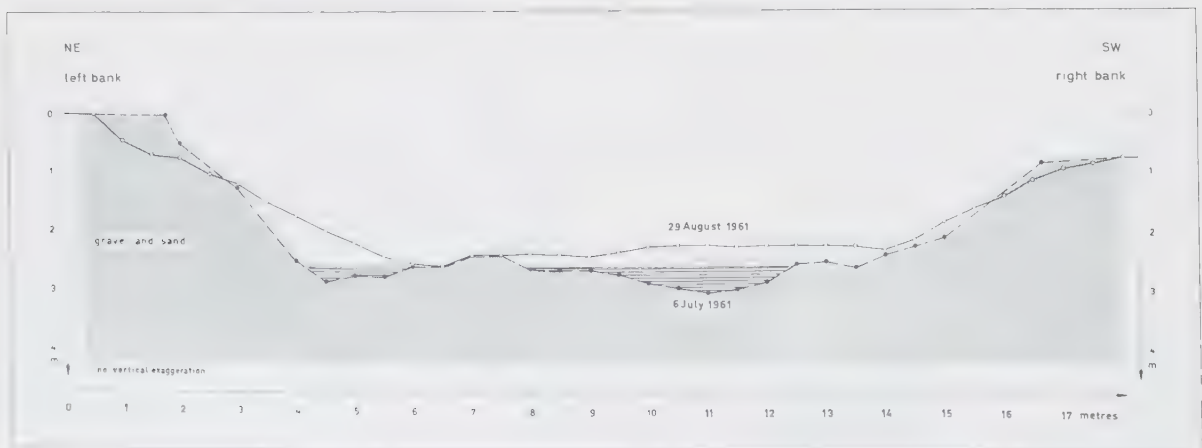


Fig. 46: Between Lake basin, recent changes in the course of an influent river.

Fig. 47: Changes in the cross-section of a river flowing from White Glacier into Between Lake, summer 1961. (Site of cross-section see Fig. 46)



Changes in sedimentation: The abrupt and frequent variations in grain size both horizontally and vertically, and the redeposition and disturbance of deposits in periodically draining ice-dammed lakes are discussed in the next sections.

3) Deposits in Lake Basins

a) Dissolved and Suspended Material in Between Lake

Sampling and analysing methods: Surface water samples were collected in polyethylene bottles of 1000 and 500 cm³ capacity. To obtain samples from different depths a self-closing water container (2 litres) was used from a dinghy. The sampling sites in Between Lake are indicated in Figures 13b and 16.

The amount and composition of dissolved substances was determined at the Kantonales Laboratorium Zürich, Switzerland. The suspended material was analysed for grain size with a micro-sedimentation-analysis apparatus (Gessner, 1963). Further remarks on the analyses of suspended and dissolved material are to be found in Chapter C.

The number of water samples taken was too small to allow an evaluation of the average suspension and solution load of Between Lake or the variations during the summer and from year to year. But the available data give an order of magnitude.

Depth samples and surface samples: On 25 July 1963 depth and surface samples were collected near the middle of Between Lake. No significant differences in total amounts or composition of dissolved substances were found (Tables 8, 9). But there seems to have been a slight stratification of the suspended material, although a strong wind was mixing the lake waters prior to and during the time of the sampling and although the lake was then draining. On a calm day there would probably be more marked differences between depth and surface samples. This was indirectly suggested by the fact that after the lake had completely drained the mud-cover was limited mainly to the lower parts of the lake basin. Of course, these lower parts were submerged for a longer time, but nevertheless the clear boundary between a substantial layer of mud covering the lower sections of the lake basin slopes and the slight or even non-existent mud-cover of the higher parts could mean that - in addition to the possible washing down of the mud from the higher to lower parts of the lake basin especially during the emptying phase of the lake - the suspended material brought into the lake basin by the influents is concentrated mainly in the lower water layers. This would cause an increase of density at depth. There are not enough data, however, to show whether or not density currents exist at the bottom of Between Lake as has been discussed for glacial lakes elsewhere (Johnston, 1922, p. 381; Kuenen, 1951; Mathews, 1956, p. 544-546). The grain size distribution of the suspended material in the depth sample is shown in Figure 48.

It seems that in 1963 the concentration of dissolved substances in Between Lake was lower than in 1961 (Table 8). One of the reasons for this might be that in 1963 the lake contained less rainwater but more glacial meltwater than in 1961. The chemical analysis of the water shows a relatively high amount of sulphates in Between Lake, although much less than in the landlocked Colour Lake (Table 9).

Inflow and outflow water: At the beginning of the drainage of Between Lake the overflow water contained a very small amount of suspended material, much less than in the influents, which demonstrates the function of such ice-dammed lakes as accumulation basins. At the end of the drainage, however, the effluent water contained a larger amount of suspended matter and a higher percentage of coarser grains than in the early stages (Table 8, Fig. 48). This probably means that a substantial amount of the sediment deposited during the filling process is removed again by the last lake water draining out and also by the river waters flowing across the empty basin. Several authors have noted similar effects in lakes with sub-glacial drainage (Stone, 1955, p. 28; Reid and Clayton, 1963; Karlstrom, 1965, p. 137, 139). The solution load of the outflowing water was also observed to increase with progressing drainage, at least in 1961 (Table 8).

b) Lake Deposits

Sampling and analysing methods: It would be ideal to have a set of samples distributed at random over the whole lake basin. But it was not possible to obtain this due to the inaccessibi-

Table 8: Some data on suspension and solution load, Between Lake, 1961 and 1963

Date	Sampling Site	Dissolved Substances mg/l		Suspended Material mg/l	Remarks
1963 26 July	Inlet from White Glacier	34	85.3	535	
1961 6 Aug.	"	-	-	1280	
1963 25 July	Middle of lake surface	65	89.8	80	{ windy; 1½ days after start of lake drainage.
	6 m depth	-	-	170	
	11 m depth (=approx. 2 m above ground)	63	91.5	165	
1961 23 July	Near outlet	98	88.8	45	3 hours after start of lake overflow. during drainage of lake.
28 "	"	-	-	35	
30 "	"	192	87.5	35	
5 Aug.	"	-	-	670	
1963 July	Near outlet	-	-	very little	before lake overflow.
23 "	"	65	89.2	42	3 hours after start of lake overflow.
25 "	"	-	-	80	during drainage of lake.
27 "	"	-	-	360	during last phase of lake drainage.

Table 9: Analysis of water samples from Colour Lake, Between Lake and Between River, summer 1963⁽¹⁾

	Colour Lake (land-locked)		Between Lake (ice-dammed)		Between River (glacial)	
Sampling Site	Surface	At Depth 23 m below sur- face, just above lake bottom	Surface	At Depth 11 m below sur- face, 2 m above lake bottom	Surface nr. shore	Surface nr. shore
	nr.outlet	middle of lake	nr.outlet	middle of lake	400 m from snout	nr.snout
Date of Sampling	26 Aug.	25 Aug. (*26 Aug.)	23 July, 2000	25 July, 0400	24 July, 1730	13 Aug., 0430
Water tempera- ture (2) (°C)	1.7	5.8	2.2	2.3	0.5	-
pH ⁽³⁾	4.3	3.1*	-	-	6.6	6.0
O2 (mg/l)	-	15.4 ⁽⁴⁾	- (5)	-	-	-
SO4 (mg/l)	-	214.0	52.3	40.5	35.8	109.7
NH3 (mg/l)	3.8	9.5*	0.05	0.06	0.05	0.14
NO2 (mg/l)	-	< 0.005	< 0.005	< 0.005	< 0.005	< 0.005
NO3 (mg/l)	-	< 0.25	< 0.25	< 0.25	< 0.25	< 0.25
PO4 (mg/l)	< 0.04	≤ 0.02	< 0.01	< 0.01	0.02	~ 0.01
Cl (mg/l)	< 0.01	0.7*	-	< 0.4	< 0.4	0.4
Ca (mg/l)	-	-	-	11.6	28.8	~30.0
Mg (mg/l)	-	-	-	29.5	17.5	19.5
Mn (mg/l)	-	-	-	< 0.1	< 0.1	0.32
Fe (mg/l)	-	~ 0.02	~ 0.02	~ 0.01	~ 0.02	~ 0.02
Hardness as CaCO3 (mg/l)	-	-	-	-	-	-
Carbonate	-	7.5	35	25	12.5	10
Total	-	204	38	40	41	102
Total amount dis- solved substances (mg/l)(dry residue)	-	344	65	63	63	178
Non-organic % of total	-	90.9	89.2	90.5	96.8	97.8
Remarks	cold and windy		start of lake drainage	windy and lake draining	flood discharge (35 m³/sec.) 1 day after start of Between Lake drainage	small dis- charge (1 m³/sec.) end of season

(1) Analyses made by Kantonales Laboratorium Zürich, Switzerland (Chapter C 3a),

(2) Measured with thermistors, (3) Determined in field with test liquids and verified in Zürich,

(4) Average of 3 samples, (5) No analysis.

lity of certain areas. However, an attempt was made to collect samples at various distances from the influents and at various elevations in the lake basins.

In Between Lake (Figs 13, 16) and Ice Cave Lake some 30 sites for measuring sedimentation were installed by laying down bases ($\frac{1}{2}$ to 1 m²) of plastic sheet, aluminum foil, cloth, gypsum, and, in addition, some 15 tins each with 100 to 300 cm² bottom area. The accumulation at these test sites should indicate the amount and kind of sedimentation during one summer, i.e. during one filling and one emptying of an ice-dammed lake. In addition, some 20 sand and silt probes were collected at different distances from the mouths of the various influents, and in connection with some temperature measurements, 3 soil profiles were dug in the active layer of the Between Lake basin from which 15 further samples were taken. It was planned to dig another pit down into the permafrost but in 1963 this was not possible because of the wet upper soil layers which, due to poor weather, did not dry out until the end of the summer. For future research a deep pit would be a worthwhile project; it would help to elucidate the history of Between Lake and therefore that of the damming glaciers.

The sedimentation investigations in lakes like Between Lake, Crusoe-Baby Lake, etc., could probably be improved by using some kind of a "sedimentation tower" (Fig. 49) or a stairlike arrangement of sampling sites, for instance at a central island or on an accessible slope (Fig. 50), so that at the same distance from influents and effluents but at varying heights over the lake bottom a series of comparable samples would be obtained.

It was only possible to collect bottom samples from two positions in Phantom Lake. They were obtained by using a bottom sampler through a hole in the ice cover, from a depth of about 200 m (Fig. 34). Unfortunately several attempts, at the same place, to lift an undisturbed sediment core to the surface, failed. Bottom samples were also collected in Colour Lake, from a dinghy.

The lake sediments were analysed mainly for grain size at the Geography Department of the University of Zürich. The whole probe was put wet through a sieving machine with a standardized set of sieves (Dyn 4188, mesh width: 4000, 2000, 1000, 500, 250, 125, 90, 63 microns). The fines (smaller than 63 microns) were further analysed down to about 1.5 microns by the hydrometer method after dispersal by a small amount of Calgon. Samples too small to be treated by these methods were analysed in the same way as the suspended material. Comparative tests were conducted on the same sample by all methods used; the results were found to agree fairly well. For the determination of the weight of the total summer accumulation, the samples were first dried in an oven at 105°C. Two samples were taken to Zürich undisturbed so that the internal structure could be investigated. Some thin-sections and microphotographs have been prepared by Dr. M. Müller-Vonmoos, Laboratory for Hydraulic Research and Soil Mechanics, Federal Institute of Technology (ETH), Zürich, Switzerland.

Amount of sedimentation: The average amount (dry weight) of summer sedimentation in Between Lake calculated from some 30 samples collected in 1961 and 1963 was about 130 grams per 100 cm². This corresponds to an average mud layer of some 0.5 cm thickness and accords well with the estimated total sediment load brought into the lake basin by the influents. The sedimentation rate in the short-lived Ice Cave Lake was about twice as high as that of Between Lake. No other sedimentation records exist for ice-dammed lakes in the Arctic, but estimates for non-arctic mountain lakes with influents of glacial meltwater in Canada and Switzerland seem to compare fairly well with the above figures (Johnston, 1922, p. 379; Collet, 1925, p. 261; Niklaus, 1967, p. 57, 64). In a glacial lake in British Columbia, Canada, Mathews (1956, p. 550-551) found a lower sedimentation rate, and for non-glacial arctic lakes in Spitzbergen the average rate (Häggblom, 1963, p. 103) is much lower than that of Between Lake.

The amount of sediment deposited during a summer varies greatly throughout the lake basin depending on the duration of submergence and proximity and character of the influents. In addition, the sediments are modified during and after deposition, through resorting by wave and current action, especially during the outflow of the lake and by occasional rains subsequent to emptying (Tarr, 1908, p. 99; 1909, p. 88-89; Karlstrom, 1965, p. 131). There was a marked difference between the amount of material that accumulated in tins, with an edge that prevented the sediment being washed off and on open sampling patches, at the same site (Fig. 51b). The amount on the sampling patches laid out on the small, more or less horizontal terraces on the slope was usually smaller than that in the nearby tins, whereas the amount deposited on patches situated in depressions at the

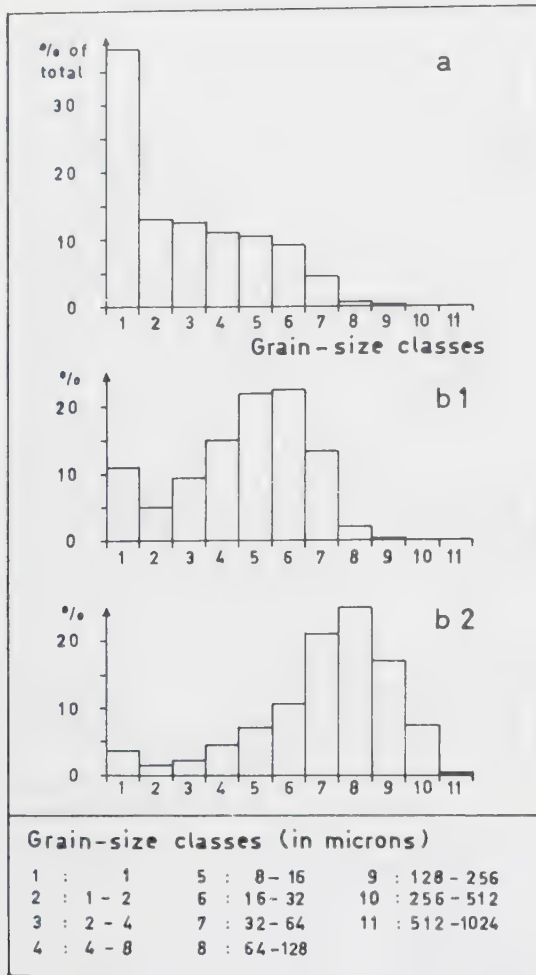


Fig. 48: Grain-size distribution of suspended material in water samples from Between Lake, summer 1961 and 1963: a) taken up at the middle of the lake from a depth of 11 m (2 m above ground), b) surface samples collected near outlet at beginning (b 1) and at end (b 2) of lake drainage.

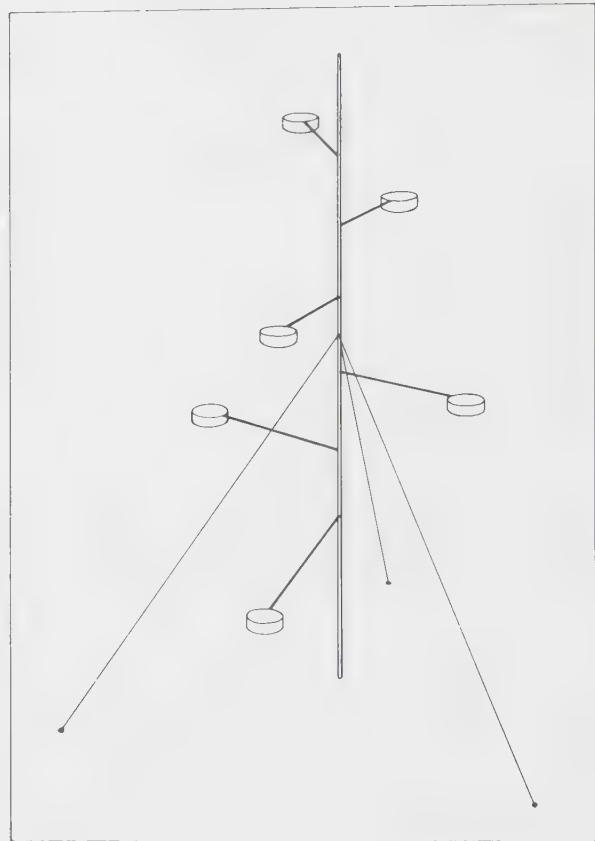


Fig. 49: Sedimentation "tower".

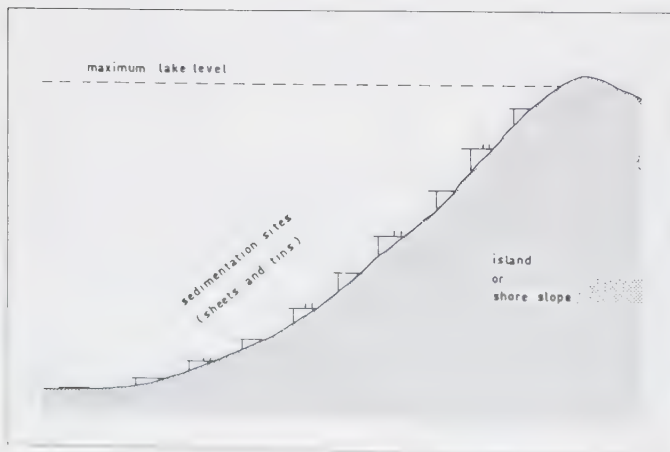


Fig. 50: Sedimentation "stairs".

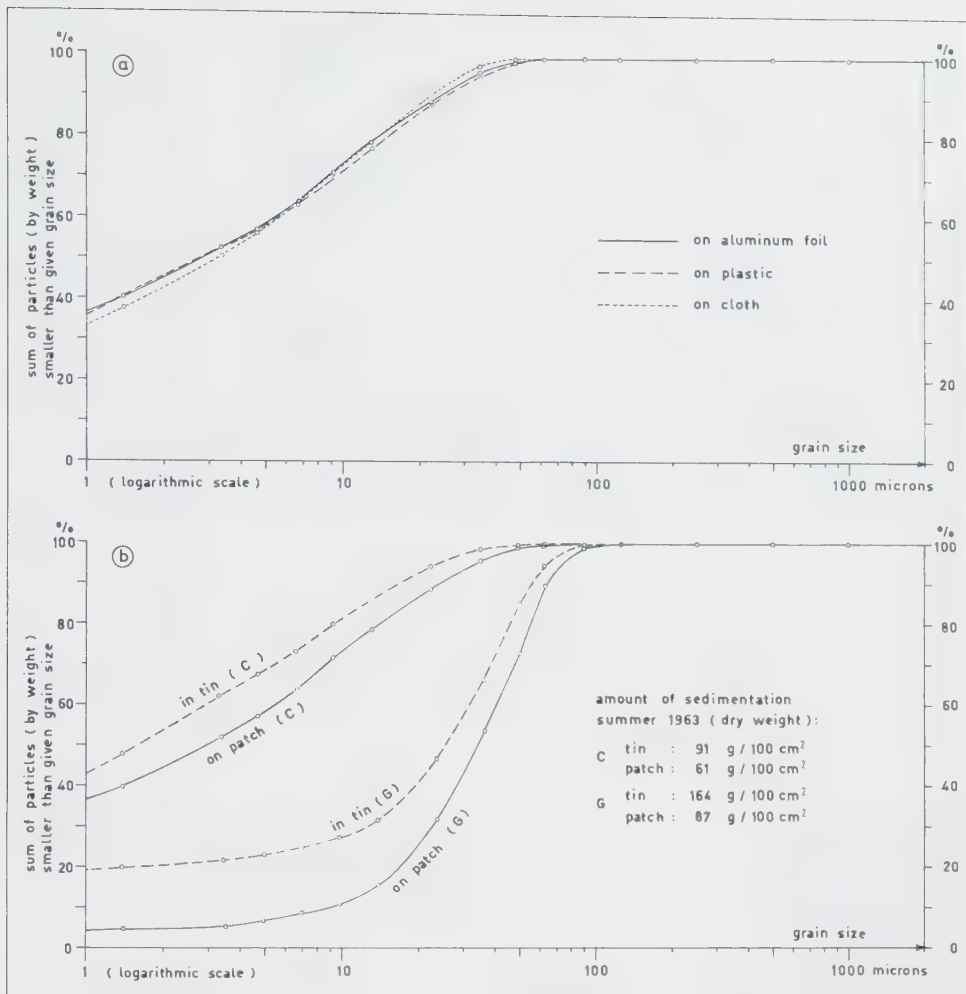


Fig. 51: Grain-size distribution in sediments deposited during summer 1963 in Between Lake: a) at site C on different test patches, b) at sites C and G in a tin and on an aluminum foil patch.

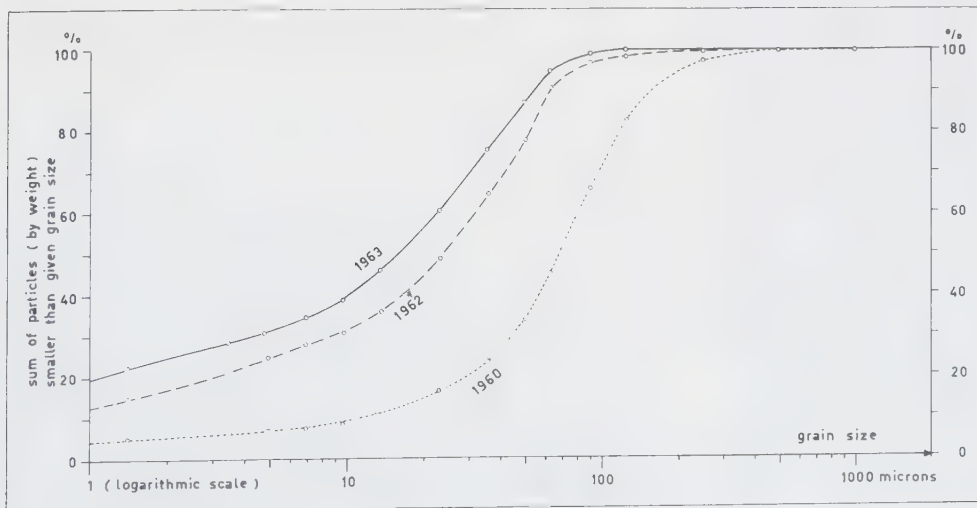


Fig. 52: Grain-size distribution in sediments deposited in Between Lake at test site A on aluminum foil during the summers 1960, 1962 and 1963.

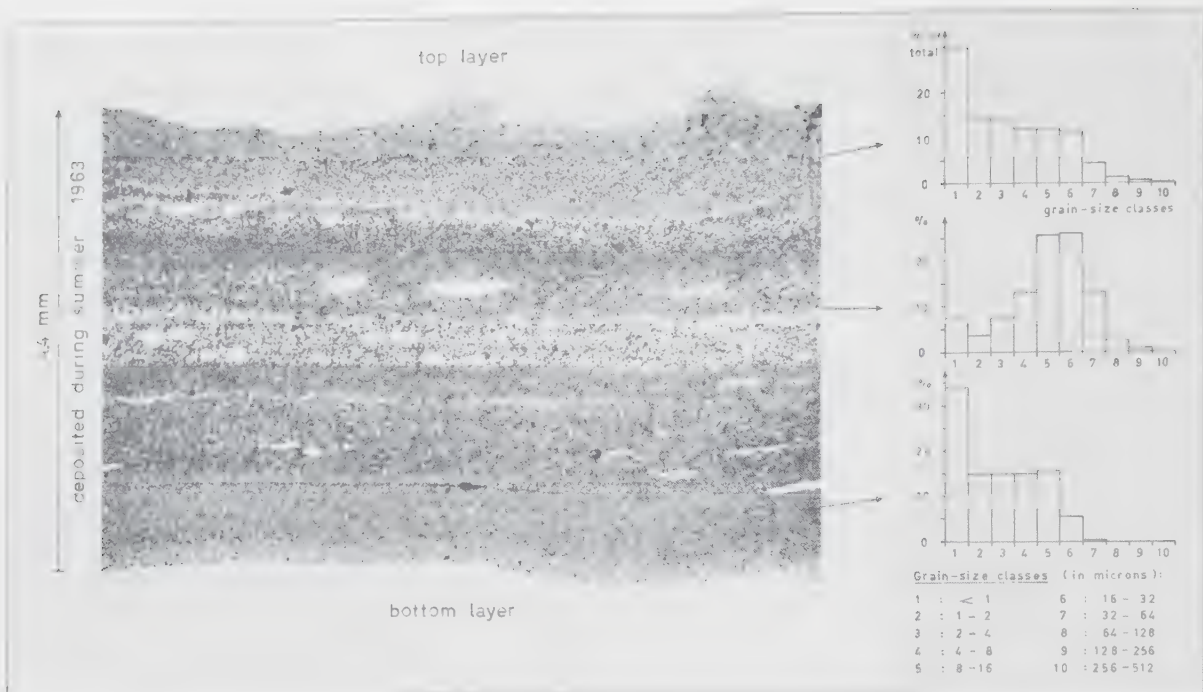


Fig. 53: Internal structure of lake sediment deposited during summer 1963 on an aluminum foil sheet (Between Lake test site A). Duration of submergence 23 days (one filling + one emptying of lake). Total thickness of dry summer deposit 4.4 mm. Microphoto of thin section + grain-size histograms of 3 sub-layers.

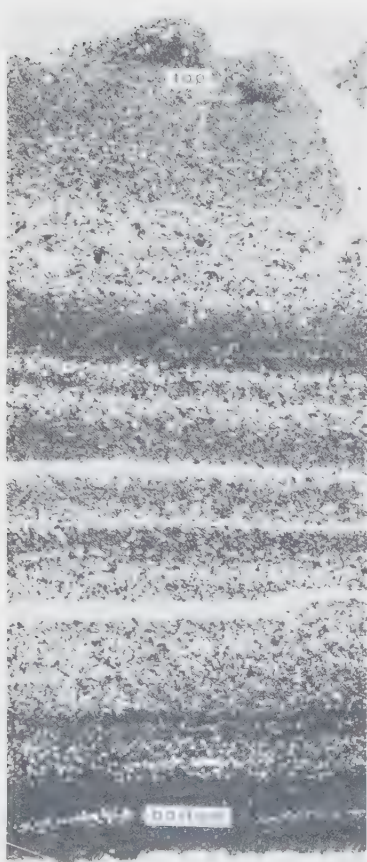
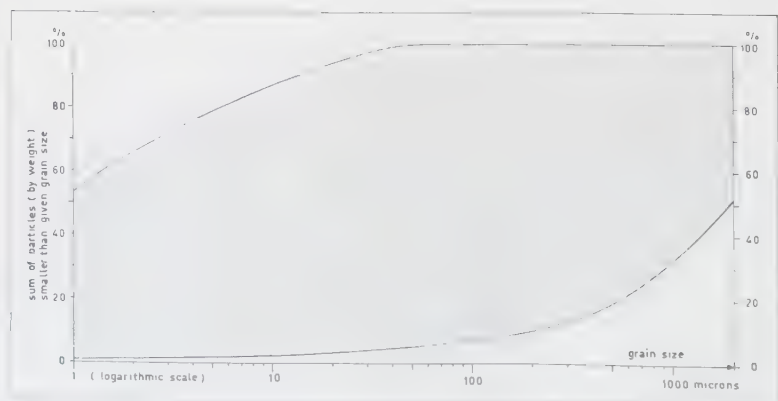


Fig. 54: Internal structure of lake sediment deposited during summer 1963 in a tin (Between Lake test site C). Duration of submergence 26-1/2 days (one filling + one emptying of lake). Total thickness of dry deposit 9.6 mm. Microphoto of thin section of dried sediment.

Fig. 55: Approximate range of grain-size distributions in sediments deposited at various sites in Between Lake, summer 1963.



bottom of the basin was larger than that in tins at the same sites. It was also observed that on the upper slopes of the Between Lake basin several plants, e.g. arctic willow, survive although submerged up to some 10 metres almost every year for a few hours or days, but that below a certain level, approximately coincident with the start of a more extensive mud cover, living plants are missing.

Grain-size investigations: The type of material used for the sampling patches did not seem to influence the particle-size distribution of the sediment that accumulated on it (Fig. 51a). But grain-size differed significantly between the sediment on the patches and that in the tins at the same site (Fig. 51b). The deposits on the patches - contrary to the protected deposits in the tins - lost a good number of the finest particles, and this mainly during the drainage of Between Lake. Such a removal of fine grains seems only to be possible in periodically emptying lakes. It is suggested, therefore, that sediments out of such lakes contain on the average a smaller amount of fine particles than sediments out of long-enduring lakes. This is corroborated by Figure 56, in which a sediment sample from Phantom Lake, an ice-dammed lake which does not drain at all or only partially, is compared with deposits collected in Between Lake, an ice-dammed lake that empties completely periodically. The comparison is limited by the scarcity of samples and differences in the petrographical environment (Fig. 55).

The grain-size distribution varies considerably in the sediment layers deposited successively at a site. The deposit sample of summer 1960, collected at test site A, was much coarser than the material that accumulated at the same site during the summer of 1962 or 1963 (Fig. 52). Within the deposit of one season, also, large variations in grain-size occur. This was observed not only in lakes where there were several fillings and emptyings during the same season, as with Ice Cave Lake in 1961, but also in lakes which fill and empty only once. Microphotographs of thin-sections were prepared from two samples that accumulated during the summer 1963 on a patch and in a tin installed in Between Lake and removed after drainage (Figs 53, 54). Distinct grain-size differences distinguish a number of sub-layers which correspond fairly well with fluctuations of the lake level and with weather and ablation data.

To obtain information on the composition of older sediment layers and on the processes that modify the lake sediments after deposition, samples were taken from the active layer on the slope of the Between Lake basin. In Figure 57 two sets of curves are distinguishable, the deeper samples show some possible influence of solifluction, the upper ones various degrees of water-sorting, probably due to the action of small rivulets and/or the fluctuating lake water. The relatively small amount of the finest particles present in the deeper soil samples could indicate - as observed in other arctic regions - some kind of sub-surface flow (Huxley and Odell, 1924, p. 212; Sharp, 1942, p. 288ff; Jenness, 1952, p. 249; Cailleux and Taylor, 1954, p. 467; Büdel, 1960, p. 65; 1961, p. 352; 1963, p. 277; Herz and Andreas, 1966, p. 262ff).

Sedimentation conditions in periodically draining ice-dammed lakes: The nature and amount of material deposited varies very much within one basin, from year to year, and from one lake basin to another. Nevertheless, there may be some common characteristics recognizable.

There are various opinions as to whether or not thinly laminated deposits can form and survive in periodically draining ice-dammed lakes. Weidick (1963, p. 112-113) states that sedimentation of "varved character" can only take place in ice-dammed lakes with outflow across thresholds, that is in lakes with continuous drainage and therefore quiet conditions for sedimentation, and not in lakes with rapid outbursts. Pettijohn (1957, p. 163, 593), in a general discussion on laminations not specifically in ice-dammed lakes, remarks that only in the absence of any bottom turbulence could such delicate laminations remain undisturbed. Larsen (1961, p. 160), on the other hand, notes that in an ice-dammed lake in southern Greenland, the 1 cm thick varves were not disturbed by the emptying of the lake except for the cutting of some grooves by the water during the drainage. Varved clay covering the bottom of an ice-dammed lake that drained catastrophically has also been observed in Iceland (Thorarinsson, 1939b, p. 237-238).

The Between Lake basin shows laminated deposits and the fine laminations were not destroyed during the emptying of the lake (Figs 53, 54). These deposits in periodically draining ice-dammed lakes are, even when of a rhythmic nature, nevertheless somewhat different from normal varves, as the causes of grain-size variations are probably much more complex and deposition is much more frequently disturbed.

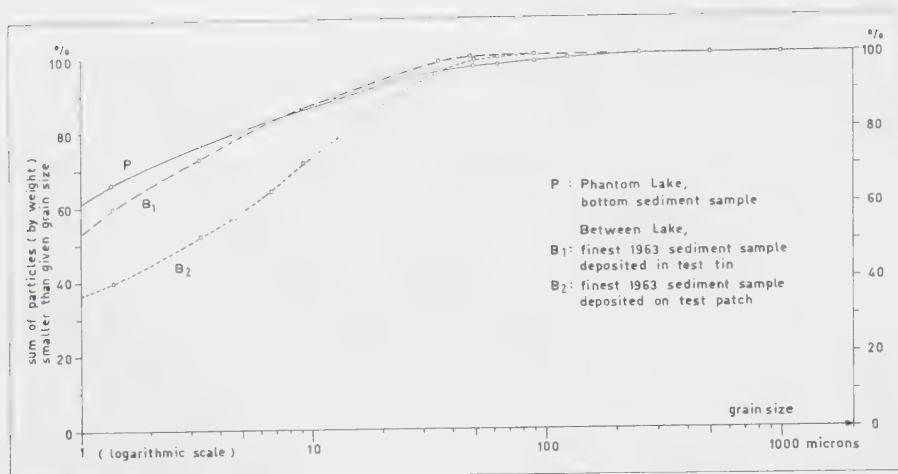


Fig. 56:
Grain-size distribution
in sediments collected
in Phantom and Between
lakes, summer 1963.

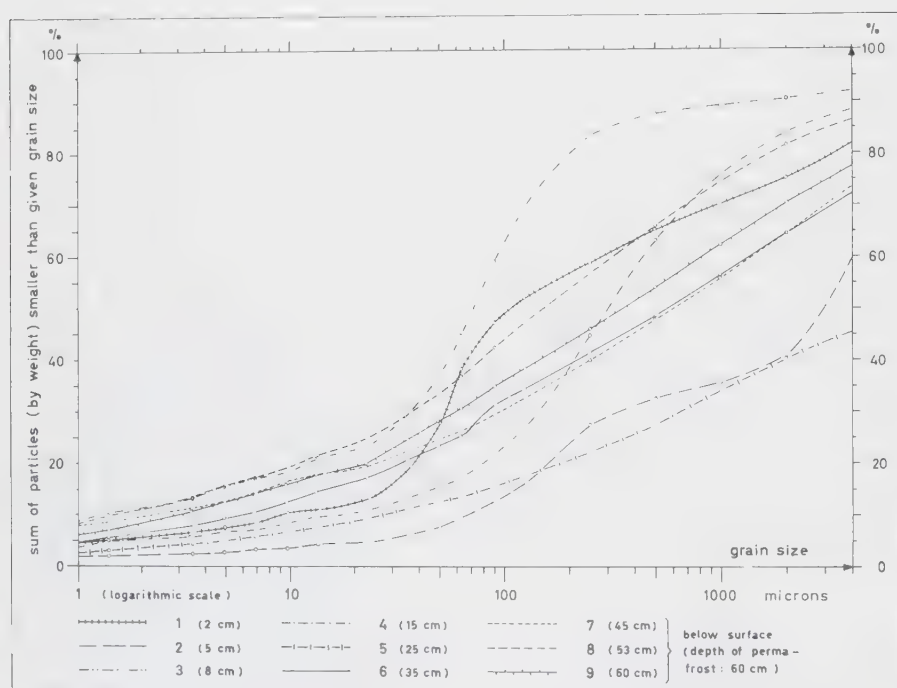


Fig. 57:
Grain-size distribution
in different soil layers.
Samples from pit 2,
Between Lake, north
shore.

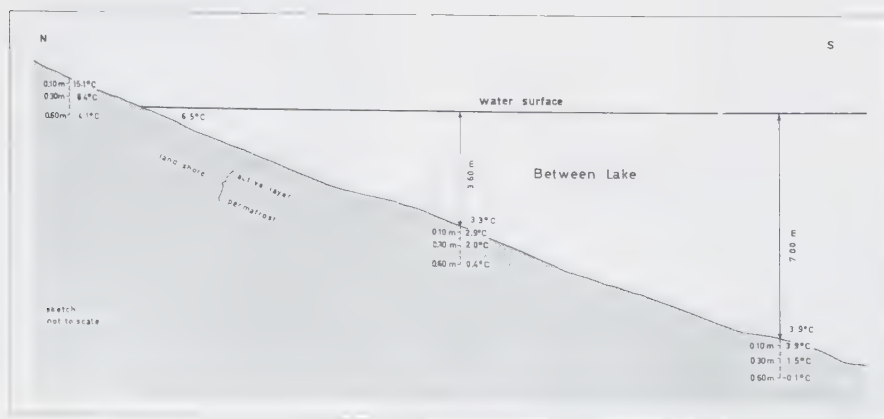


Fig. 58:
Water and soil temp-
eratures, Between
Lake basin, north
shore, 19 July 1963,
1700 (calm, sunny).

In a lake like Between Lake, which normally drains annually, there is usually only a summer deposit, whereas in a lake that stays full over winter there is also some sedimentation during the winter months even if there is no influent water, as the smallest grain fractions in suspension take several months to settle. So, in Between Lake, where the winter layer is normally missing, there is a sedimentation hiatus which is accentuated by various processes at work during the "empty season", as for example:

- mudflows, landslips, and slumping;
- melting of stranded icebergs and lake ice pieces;
- rivers flowing over the empty lake basin, eroding and depositing, producing truncation of beds, cross-bedding, ripple marks and other current effects;
- dessication of the upper sediment layers with mud cracking;
- deflation of the finest grain fractions;
- freezing and thawing of the upper some 50 cm of the sediments (the surface layers of the sediments in an enduring lake remain unfrozen);
- rain, melting snow, thawing of the active layer, produce the effects of rill erosion, solifluction, and sub-surface flow on the slopes of the empty lake basin.

Though these effects help to separate one summer deposit from the next, yet it was often difficult or sometimes impossible to distinguish annual layers as the grain-size differences between two sub-layers within the same summer deposit could be larger than between two different summer layers (Figs 52 and 53). Grain-size alone is an unsure basis on which to interpret the periodicity of glacial lake deposits.

Some of the processes that are responsible for the marked horizontal and vertical variability of the sediments deposited during a single summer in the basin of a periodically draining ice-dammed lake, are summarized here:

- washing down of the higher deposits or, at least, of their finer grain fractions by the wave action of the lowering water surface, by rain, etc., and re-deposition in lower parts of the basin;
- re-deposition of sediments that have been stirred up by water moved by wind, currents, or subaqueous slumping;
- disturbance of sediments and flushing out of certain grain-size classes during the outflow of the lake water, especially towards the end of the emptying;
- changes in the base level of erosion of influent rivers with possible shifting of discharge points and therefore complex sequences of erosion and accumulation;
- fluctuations of the discharge and sediment load of influent rivers;
- melting out and deposition of morainic material from icebergs;
- disturbances by grounded icebergs and lake ice pieces.

Consequently in such lake deposits there are seldom persistent layers, the grain-size distribution may change gradually or abruptly in a vertical or horizontal direction, small-scale cross-bedding and graded bedding (normal and reversed) can be found side by side or one above the other, although usually belonging to two different depositional environments (Pettijohn, 1957, p. 164).

Much has been written about short-term rhythmites or varve variations (Klebensberg, 1948, p. 319; Woldstedt, 1954, p. 154; Charlesworth, 1957, p. 1154-1155; Flint, 1957, p. 293, 297; Leopold, Wolman and Miller, 1964, p. 397-398), however, most of the knowledge is based on the study of fossil (mostly late Pleistocene) glacial lake sediments and not on observation of actual sedimentation processes.

4) Thermal Effects of Ice-dammed Lakes

Only sporadic, unsystematic temperature measurements could be undertaken, nevertheless, they may be of some interest, as relatively few data exist on the temperature régime and thermal effects of arctic lakes, and especially of arctic ice-dammed lakes (Sater, 1963, p. 116).

a) Water Temperature and the Thermal Classification of Lakes

The water temperatures were recorded from the shore, or from a dinghy, or through a

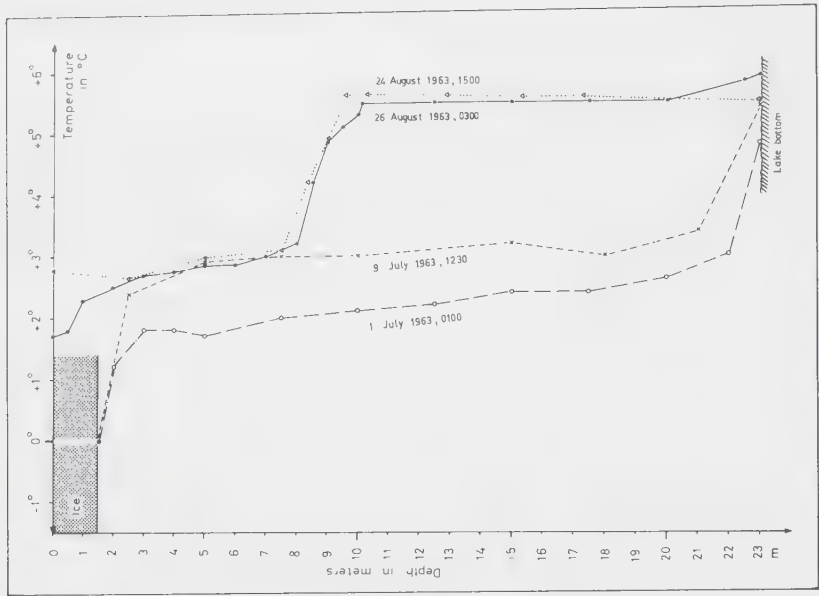


Fig. 61: Water temperatures in Colour Lake, summer 1963. For site of temperature measurements cf. Fig. 60.

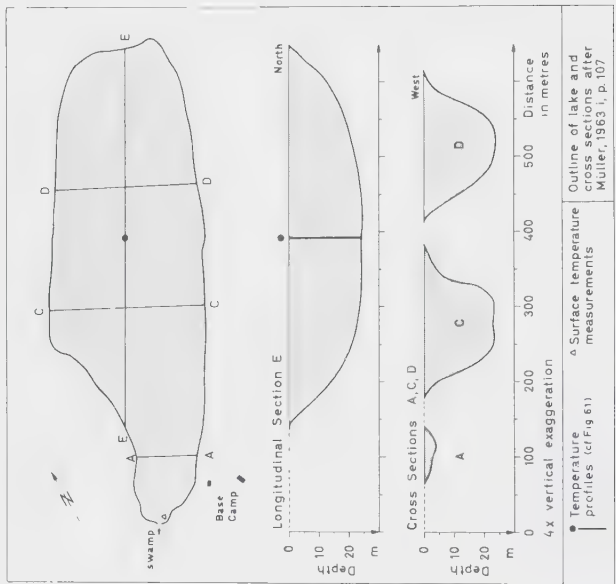


Fig. 60: Sites of temperature measurements in Colour Lake, summer 1963.

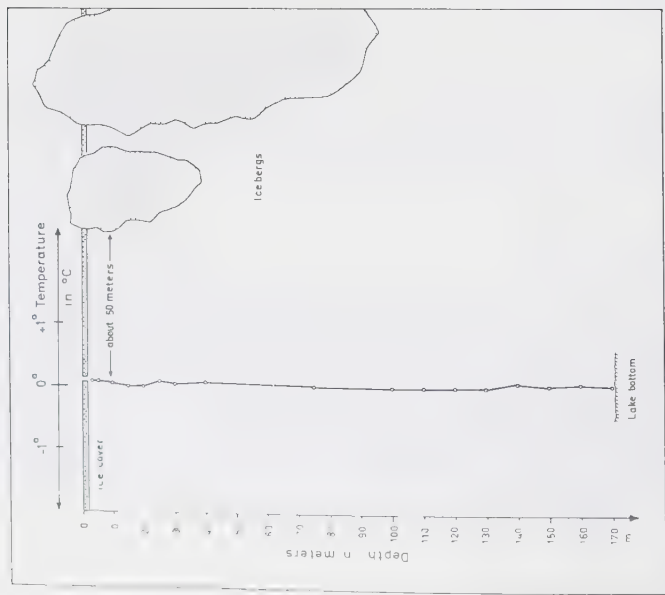


Fig. 59: Vertical water temperature profile, middle of Phantom Lake, 5 August 1963. For site of measurements cf. Fig. 34.

hole in the ice, by means of standard thermometers, a reversing thermometer, and thermistors. The thermistors and resistance bridge used were the same as those described by Müller (1961c, p. 88). As there was some doubt as to the stability of certain of the thermistors (Müller, 1963h, p. 81), an additional 0°C-calibration was made in the field, and the reversing thermometer was calibrated against the standard one.

Between Lake: Marked differences were observed between surface water temperatures near the land and close to the damming ice, especially on calm days. The water standing over the gently sloping shore of gravel, sand and silt was warmed up on several occasions above the temperature of maximum density, to as much as 6.6°C, whereas the water adjacent to the ice walls always remained below 4°. These temperature differences give rise to exchange circulations, as discussed by Thomas (1962, p. 54-56) for the Mårjelen Lake, by the Aletsch Glacier, in Switzerland.

During calm, sunny periods differences also developed locally between the surface and deeper water layers, with a relatively high-lying thermocline (Fig. 58). Such temperature stratifications were only temporary, as the water of this relatively shallow lake was frequently stirred up by winds resulting in almost isothermal conditions during which an approximate value for mean temperature and heat content of Between Lake was obtained. On 25 July 1963, during the outflow of Between Lake, the surface and depth temperatures in the middle of the lake and at the shore were close to 2°C. Measurements in the turbulent outflowing water confirmed that this was a representative mean temperature for the period of emptying. Largely because of variations in the amount of incoming radiation, which for Between Lake is the principal source of heat (Hutchinson, 1957, p. 461; Oliver, 1964, p. 75), and because of differences in the amount of ice floating in the lake and in the amount of relatively warm rain water accruing, the mean temperature of the water changes throughout the summer as well as from year to year. In 1961 and 1963 roughly the same mean temperatures for the outflowing water were recorded; in 1960 and in 1962 the water was probably a little warmer, whereas in 1964, probably due to less insolation the temperature was lower than in the preceding years. Again, in 1965, the mean water temperature must have been colder than in 1963, as the winter ice cover had not melted before the drainage started (Fig. 41).

In the light of the above temperature data, Between Lake can best be classified as a cold-monomictic lake (Löffler, 1957, p. 38) with parts of it having temporarily in certain summers a sub-polar dimictic character (Hutchinson, 1957, p. 461). Oliver (1964, p. 82) has questioned the adequacy of these terms for lakes where thermocline formation does not take place, or only locally and temporarily, and where the total mixing of the water occurs at frequent intervals. The thermal régime of such ephemeral lakes as Between Lake - and even more Ice Cave and Crusoe-Baby lakes - which exist for only a few weeks in summer is very different from that of enduring lakes. Between Lake, during a good part of its existence, is also rather shallow and with a fast changing level, which further reduces the feasibility of thermal classification (Löffler, 1957, p. 36).

Crusoe-Baby Lake: From the very few records available, it appears that the mean temperature of Crusoe-Baby Lake was - at least in 1963 - slightly lower than that of Between Lake even though a relatively smaller part of the lake bordered on ice and the emptying was lower. This rather low average temperature may be due to the shadowy lake site, the cold influents and the relatively fast filling which implies only a short time for heating by radiation.

Phantom Lake: In 1963 and 1964 Phantom Lake remained completely ice covered throughout the summer. At the beginning of August 1963 the lake ice was still some 1.90 metres thick (the upper 50 cm being candle ice), permitting the landing of a Piper Super-Cub. Lakes with persisting ice covers have been reported from other arctic regions (Katz, 1953, p. 8-9; Barnes, 1960; Roen, 1962, p. 60-62; Oliver, 1964, p. 81). Even in the warm summers of 1960 and 1962, the Phantom Lake ice cover thawed only marginally and partially broke up at only a few places. The temperature profile recorded 5 August 1963 through a hole in the ice by means of thermistors and checked with a reversing thermometer, shows almost isothermal conditions with temperatures very close to 0°C (Fig. 59). It is, of course, possible that there are minor thermoclines between the measuring points, which would become obvious only on the continuous recording of a bathythermograph. The low temperature may be due to the presence of a large number of huge icebergs, which consume almost the whole heat surplus. It was not possible to measure the temperature stratification in the more easterly part of this large lake, where fewer icebergs are present. A near 0°C temperature seems to be representative for the western half of the lake, the part choked with icebergs, and which, in warm summers, was observed to drain partially by overflowing along the

southern margin of the damming Thompson Glacier lobe. Probably even in exceptionally warm summers the Phantom Lake waters do not circulate above 4°C, so that it should be classified as cold-monomictic, and possibly in certain periods, when the lake remains completely ice and snow covered for several years, it has an amictic character (Hutchinson, 1957, p. 462). A comparison of Between Lake with Phantom Lake shows that the temperature depends decisively on the relative amount of ice that is in contact with the water.

Colour Lake: This small landlocked lake, on which the Base Camp is situated, lies 176 m a.s.l. some 8 km inland from Expedition Fiord and 2.5 km from the snouts of the White and Thompson glaciers. The size and shape are given in Figure 60. For comparative purposes, the temperatures of this lake were also investigated. The vertical temperature distribution was re-recorded on four occasions in the summer of 1963 (Fig. 61) and the conditions at the measuring site, approximately in the middle of the lake (Fig. 60), are here tabulated.

Date of measurement	Method of measurement and state of ice cover	Weather (Base Camp)
1 July 1963 about 0100	Through compact ice cover, some 1.50 m thick	Air temperature: 2.7°C Wind: calm
9 July 1963 about 0900	Through ice cover, still close to 1.50 m thick in middle but open near shore; much candle ice	Air temperature: 9.0°C (4-7 July relatively warm) Wind: some 4 m/sec.
24 Aug. 1963 about 1500	From a dinghy; still some 30% of lake covered with floating ice, near shore some new lake ice already forming	Air temperature: -3.2°C (previous 3 days also cold) Wind: some 5 m/sec. (strong winds 16-18 & 21-23 Aug)
26 Aug. 1963 about 0300	As 24 August	Air temperature: -5.1°C Wind: + calm

The relatively low surface water temperature, which always remained below 4°C, reflects the fact that in the summer of 1963 Colour Lake was never completely free of ice, whereas in 1960, 1961 and 1962 the lake was ice-free for several weeks and the surface temperatures were much higher.

The vertical temperature profiles of 1 and 9 July show a thermocline close under the ice cover, below which near isothermal conditions pertained - especially on 9 July - with a second sharp thermocline near the base and remarkably high bottom temperatures. This warm bottom water may have persisted from the previous summer and/or may result from bottom heating. The warming up of the main water body in the period 1 to 9 July must have been due largely to radiation through the ice cover, supplemented by small amounts of warmer water brought in by the land rivers and other minor heat sources such as biological processes like bacterial action (Simojoki, 1956, p. 13; Brewer, 1958b, p. 282-283). On 24 and 26 August, at the end of the summer, the lower half of the water body had warmed up to an isothermal temperature of some 5.5°C, the 4°C water was almost missing, and a pronounced thermocline at about 8.5 m separated the upper cold water from the lower warm mass. The strong winds of 16 to 18 and 21 to 23 August may have influenced the position of the thermocline; whereas the low temperatures and the wind of 24 to 26 August had a cooling effect only down to about 2.5 m. On 24 and 26 August the lake did not appear to have a stable stratification from the vertical temperature distribution. This could mean that a chemically controlled density stratification was superimposed on the temperature controlled one, i.e. that a higher content of dissolved substances in the lower water layers increased the density. There are no fish in the lake, and the low pH-value together with the preliminary analysis of surface and bottom water indicate a much higher concentration of dissolved substances than in Between Lake (Tables 8 and 9), suggesting that Colour Lake is a meromictic lake. The situation in an over-deepened, glacier-scoured trough (Robitaille, 1961, p. 175; Müller, 1963i, p. 108) with very little inflowing water and only a shallow outlet favours a meromixis. Further investigations are necessary to decide whether or not plant decay and diffusion of solutes out of the mud increase the differences in chemical composition of the surface and bottom water. The high content of sulphates in Colour Lake is due to the proximity of an evaporite diapir. There is also the possibility of thermal and chemical influence from warm, saline springs, as some exist only about 1300 m away from the lake but on the opposite side of Gypsum Hill, however, the water differs in chemical composition from that of the lake (Table 9 and Beschel, 1963a, p. 185).

The ice-dammed lakes investigated in the Expedition Area are of a cold-monomictic (or even amictic) character, whereas the landlocked Colour Lake could be called - although winter temperature records are lacking - a sub-polar dimictic lake with a possible meromixis. Hutchinson (1957, p. 461) states that the cold-monomictic lakes owe their thermal character primarily to the adjacency of ice rather than to the high latitude.

b) The Effect of the Lake Water on the Glacier Ice

The influence of the Between Lake water on the temperature régime of the upper ice layers of the glacier was examined, paying special attention to the speed of melting of ice under water as compared with ice exposed to air.

Three sets of thermistors were frozen into holes drilled in the ice at depths of 20, 50, 80 and 150 cm, with one other thermistor fixed about 20 cm above the ice surface. Two of these thermistor sets became submerged as the water level of Between Lake rose, but the third was situated above maximum water level (Fig. 65). An improvised device for measuring the surface lowering of the ice (Fig. 62) under water was installed and worked well until the wire from the submerged pole to the shore broke; after that the amount of surface lowering could only be determined indirectly from the temperature records of the thermistors which melted out under water one after the other. The measurements are too few and too widely spaced for anything but a tentative interpretation.

Figure 63 shows that the ice under water melted faster than that exposed to the air. This agrees with the findings of Mercanton (1955, p. 271-272) on the influence of the artificial Grimsel reservoir on the submerged tongue of the Unteraar Glacier, a temperate valley glacier in the Swiss Alps. Recent observations, reported by Kasser (1965, p. 243, 245; 1967, p. 226, 229), on the variations of the tongues of the Unteraar and Oberaar glaciers again show that submergence by the waters of the reservoirs accelerated melting. Okko (1956, p. 103) and Sissons (1958b, p. 165) also cite the more rapid melting of ice in contact with lake water. The deepening and enlarging of glacial ponds by the melting activity of standing water was noted by Agassiz (1841, p. 192-193), Renaud (1936, p. 18, 27), Sharp (1947, p. 44; 1949, p. 308, 312-313) and Holmes (1955, p. 46). Typical forms produced on the ice surface by subaqueous melting are shown in Figure 64. The shorelines "cut" into the damming glacier walls are obvious signs of melting by lake water, especially by the relatively warm surface water. The effect of moving water on ice is even greater than that of standing water. Near and at the outlet of Between Lake an accelerated lowering of the submerged ice surface was observed as soon as the lake started to drain (Fig. 63).

There are some indications that the temperatures in the ice submerged by water were some 0.5°C higher than those recorded at comparable depths in the ice above the maximum water level (Fig. 65). The thermistors recorded increasingly higher temperatures due to the thinning of the overlying ice by melting and the downward penetration of heat. At first the gradient of this warming up was about the same at both the drowned sites and at the non-submerged one. But, as soon as the thermistors came close to the surface, the warming appeared to proceed faster at the sites covered by water. Water temperatures of close to 2°C were recorded before and during the lake outflow by both the thermistors fixed above the ice surface and those which melted out of the ice and floated in the water.

c) Thermal Effects of the Lake Water on the Ground

Echo depth sounding and observations during bottom sampling suggest that the Phantom Lake bottom is soft for at least some metres. The downward extent of the thawed zone beneath this large ice-dammed lake could not be determined. In studies on the thermal effect of lakes and oceans it has been maintained that even under relatively small shallow lakes there exists a thick thawed zone between the water and the permafrost and that beneath large water bodies there is no permafrost at all (Werenskiöld, 1953; Brewer, 1955; 1958b, p. 282; Lachenbruch, 1957b; Grigor'ev, 1959a; Legget et al., 1961, p. 963-964; Tsvetkova, 1961; Johnston and Brown, 1961; 1964; 1965; Brown, 1963; Hussey and Michelson, 1966, p. 172). The latter may be the case for Phantom Lake, although the water temperature, at least of the western part, is very close to 0°C. However, the extent of the thawed zone and the location of the permafrost table under and about the lake are not yet known.

Even shallow lakes that freeze completely to the ground in winter exert some influence on

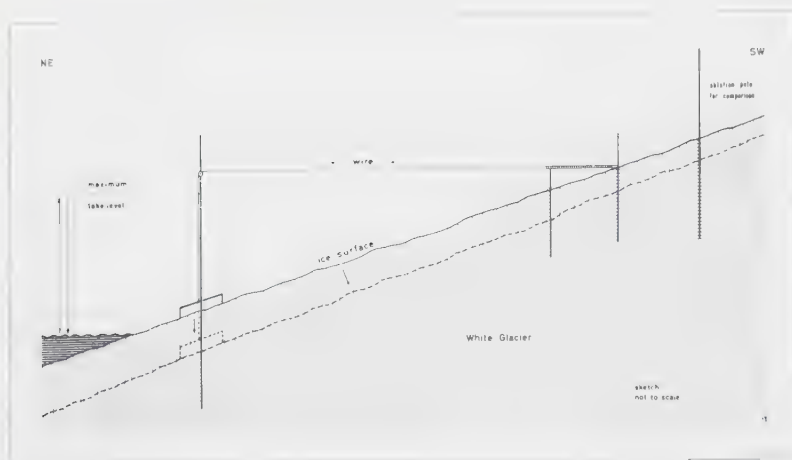


Fig. 62: Device to measure ice melting under water (Between Lake, summer 1963).

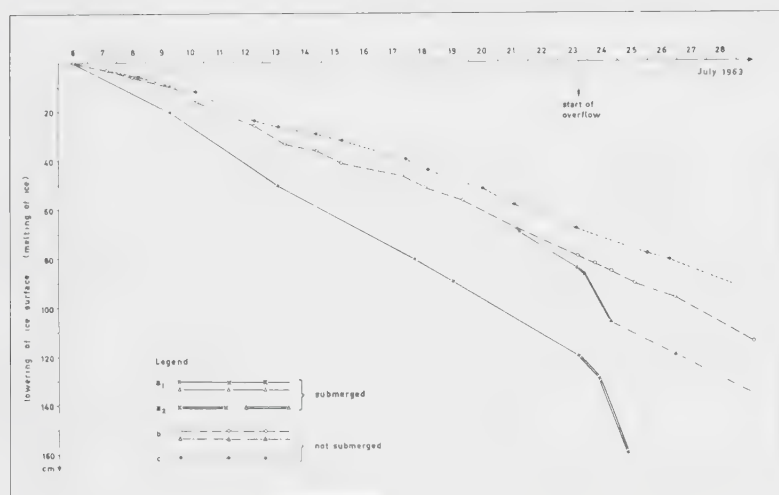


Fig. 63: Comparison of ice melting at submerged and unsubmerged sites by Between Lake water, July 1963: a₁) submerged ice surface (inclined some 5°), before lake overflowed (quiet water), a₂) ice surface (inclined some 5°), during outflow of lake (moving water), b) unsubmerged ice surface (inclined some 5°), c) unsubmerged ice surface (horizontal), Lower Ice Station, average out of 5 stakes.

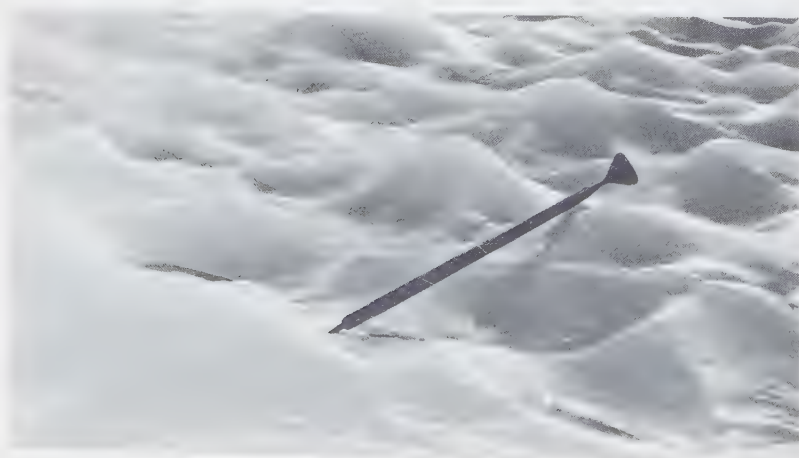


Fig. 64: Meltforms on ice surface, summer 1963.

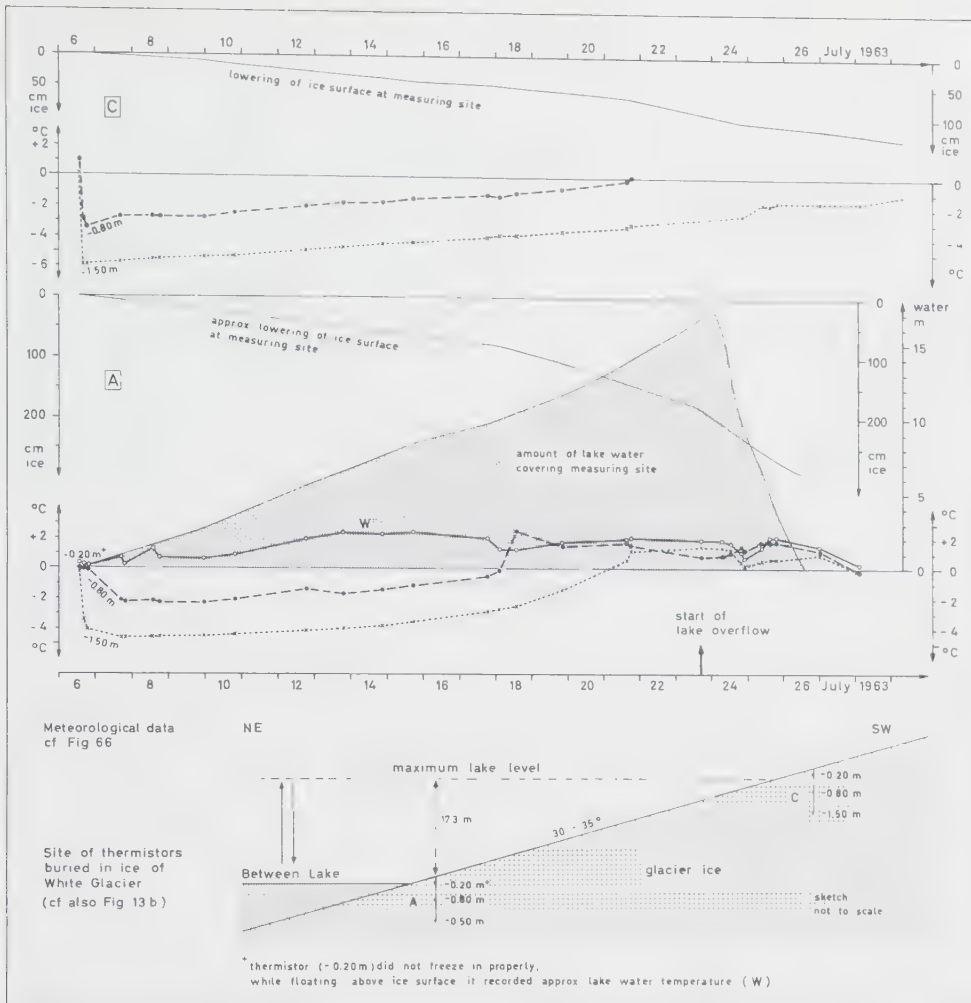


Fig. 65: Ice temperatures during filling and emptying of Between Lake, summer 1963:
 C) site not submerged by lake water, A) submerged site.

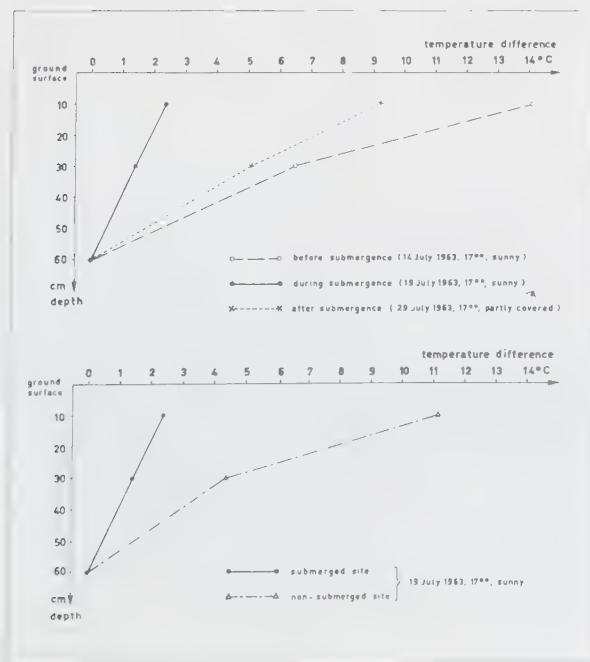


Fig. 67: Temperature differences between various soil layers (Between Lake basin).

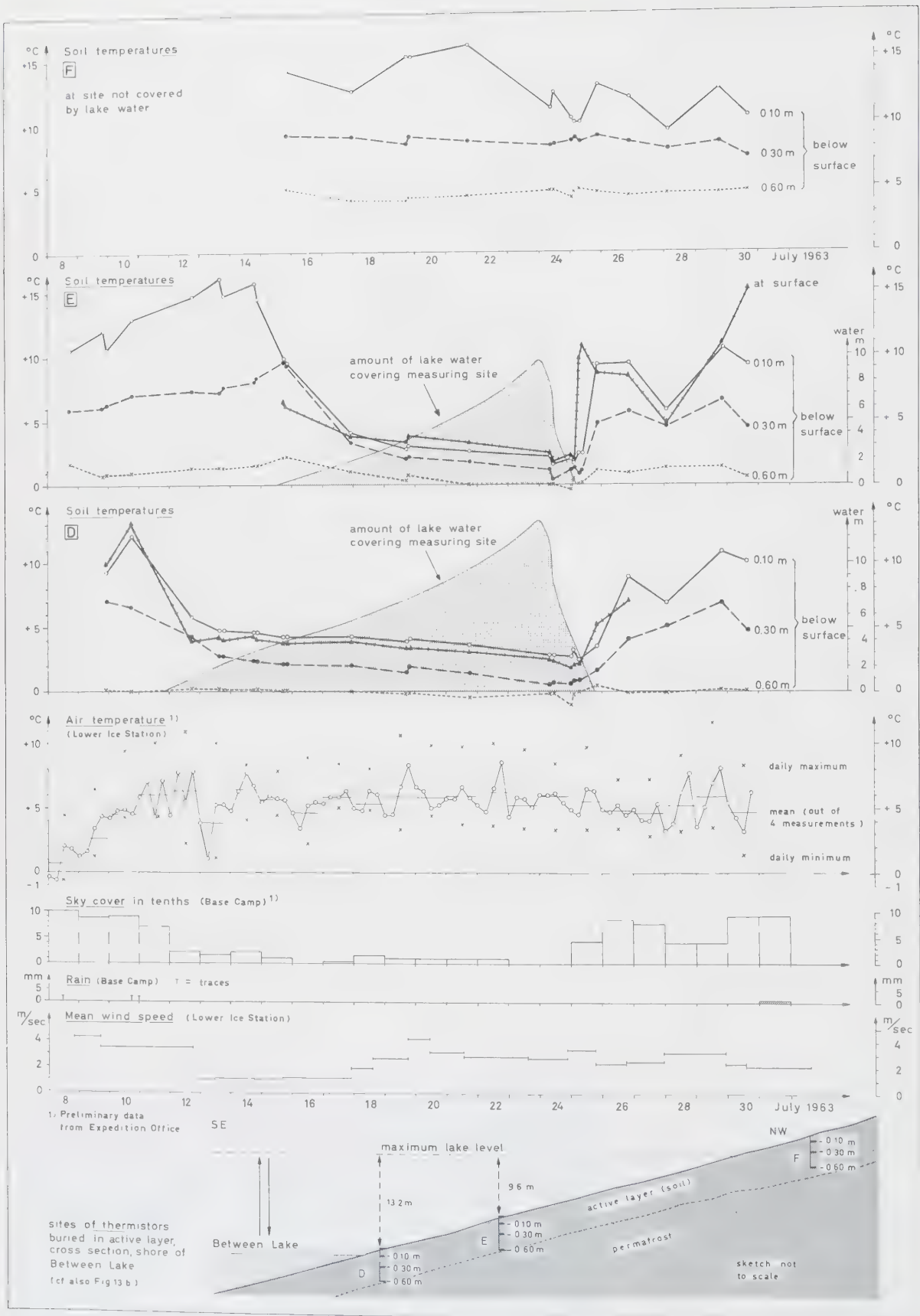


Fig. 66: Soil temperatures during filling and emptying of Between Lake and comparison with meteorological data.

the permafrost beneath (Legget et al., 1961, p. 964). So it is possible that the relatively cold, periodically draining ice-dammed lakes with their fast fluctuating levels and with the basin often empty in winter, may have some influence on the thermal régime of the ground. To measure the effect of a short submergence on the upper soil layers, 3 groups of thermistors were inserted in the active zone of the gentle slope of Between Lake (Fig. 66). The two lower sets, each with one element measuring the temperature just above the ground surface, were submerged for some days when the Between Lake level rose; they appeared again when the lake drained. The highest set, above the maximum water level remained dry. The scope of this investigation was very limited. No drilling or digging in the permafrost was possible, therefore, the thermistors could only be inserted in the upper 60 cm of the ground. The submergence was short, and only sporadic readings could be taken as no thermograph was available, so that minor temperature fluctuations were unobserved. The thermal conductivity (Lachenbruch, 1957a) could not be measured. There is also some disturbance of the structure and thermal régime of the active layer during the installation of the thermistors.

The moderating effect of the water was obvious. The ground temperature fluctuations were much smaller, especially in the 10 cm and 30 cm layers (Fig. 66), and the differences between the temperatures at the various depths were also much reduced during the period of submergence (Fig. 67).

The interval between the measurements was too long for accurate assessment, however, the phase lag of short-term diurnal, fluctuations, which in non-submerged ground can be several hours at a depth of some 50 cm (Cook, 1958, p. 7; Weedfall, 1963, p. 185), seems to be much reduced by submergence. The warm shore water very quickly influenced the temperature, even at the 60 cm level (Fig. 66, site E). The thermal conductivity and the heat storage capacity of the ground changed as soon as it was soaked by the lake water (Wundt, 1940, p. 405; Sharp, 1942, p. 289; Geiger, 1957, cited in Weedfall, 1963, p. 184; Legget et al., 1961, p. 964-965; Cook and Raiche, 1962, p. 65, 76; Herz and Andreas, 1966, p. 272).

Cook (1958, p. 8-9) stresses the fact that in summer, solar radiation has more effect on soil temperatures than does air temperature, referring to non-submerged areas. It is evident from Figures 58 and 66 that during submergence, especially when the water column reached a height of more than 2 m, the direct influence of radiation on the ground was only very slight and the soil temperature depended mainly on the water temperature. There was only a small difference between the bottom water temperature and the soil temperature at a depth of 10 cm. With the increasing height of the water column the bottom water temperature, and therefore also the soil temperature, decreased. In summer the temperatures of the near-surface soil layers are higher and less stable at the non-submerged sites than at places covered by the relatively cold waters; in winter the reverse would be true. But in lake basins that drain late in the summer and do not have time to dry out, the cold can penetrate more readily as the thermal conductivity is increased in the wet ground as compared with adjacent dry areas.

The marked cooling on 24/25 August, just before the end of the period of submergence (Fig. 66), is difficult to explain, perhaps it was due to the strong winds at that time, mixing the whole water mass and bringing 2°C water even to the shore.

When considering the possibility of changes to the permafrost table during submergence, it is noteworthy that only small temperature fluctuations took place at the 60 cm depth, which at sites D and E is the layer just above the permafrost table. Part of the heat that penetrated down to the 60 cm level at these sites, just after the start of submergence, may have instigated some melt in the frozen ground beneath, but, fairly soon, the soil temperatures became more or less stable at about 0°C. This indicates, perhaps, that the heat used for melting was compensated for by the latent heat of fusion liberated during freezing, the "zero curtain" in Muller (1947, p. 17). The negative temperatures recorded towards the end of the transgression do not necessarily mean that the thermistors were frozen in at that time, as temperatures lower than 0°C are required to induce freezing in soil (Cook and Raiche, 1962, p. 65, 75-76). So, the submergence by the waters of Between Lake might have, at most, influenced the thermal distribution within the upper layers of the permafrost, but the duration was most probably too short to cause a change in the location of the permafrost table. When Between Lake remains full over winter, however, as in 1964/65, a marked extension of the thawed zone beneath the lake basin must occur.



Fig. 68: Double-walled channel (foreground) and single-walled channel (background) at left margin of Crusoe Glacier, cut by Accident River into frozen moraine and gravel (23 June 1963).



Fig. 69: Double-walled channel at left margin of Crusoe Glacier, cut by Accident River through broad spur of frozen bedrock (shale) with thick layers of ground ice (22 July 1961).

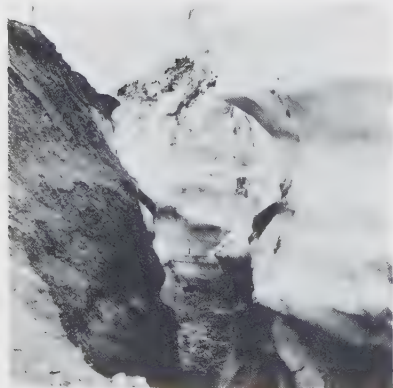


Fig. 70: Single-walled channel at left margin of Crusoe Glacier cut by Accident River. Ice axe for scale (3 August 1961).

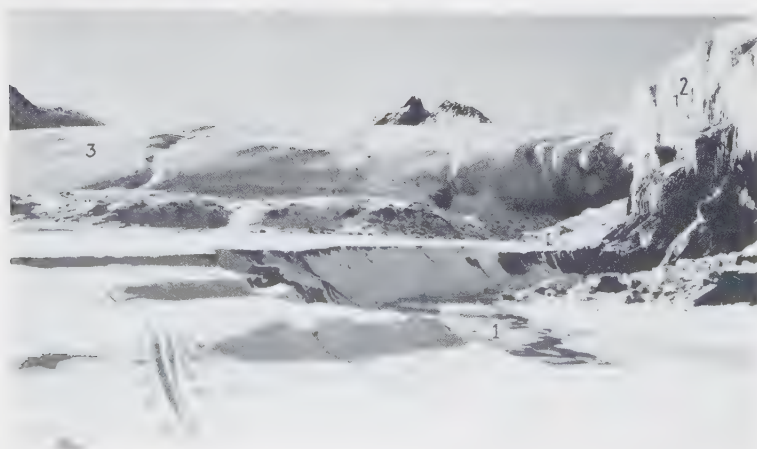


Fig. 71: Erosion terraces cut by Between River (1) when dammed by ice fall from Thompson Glacier (2). In background tongue of White Glacier (3) with terminal moraine (28 May 1961)

Chapter C

INVESTIGATIONS OF MARGINAL AND PROGLACIAL MELTWATER RIVERS

As previously mentioned, practically all the meltwater from cold type glaciers drains marginally and supraglacially. Some observations were made on the geomorphical effect of these meltwater streams.

1) Marginal and Sub-Marginal Channels of existing Valley Glaciers

Much has already been written about glacial drainage channels. Most of these papers are descriptions and interpretations of fossil features mainly in Scandinavia, the British Isles and North America. Relatively few observations on channels in the process of formation along present day glaciers are known, and of these the majority is from temperate glaciers in arctic and sub-arctic regions and only very few from cold type glaciers in the Arctic.

Position and morphology of glacial drainage channels and the speed and manner of their formation depends on various environmental factors, both climatically determined and local, e.g.: temperature, dynamics, regimen, thickness, surface gradient of the glacier; channel elevation in relation to the equilibrium line; grade, regimen, volume, load, temperature of the marginal river; local topography (preglacial relief included), of special importance being the steepness of the valley slope, the occurrence of spurs, ravines, cols; local lithology and geological structure; permafrost conditions; existence of lateral moraines; presence or absence of ice-dammed lakes (and their drainage regimen); and duration of channel occupation.

As many of these factors change from place to place there is a great variety of glacial erosion features. This diversity of forms and several of the above determining factors have been discussed by various authors, especially by Ives and Kirby (1964, p. 917) and also by Flint (1957, p. 168), Derbyshire (1958, p. 191), Ives (1959, p. 58). The observations along the margins of the sub-polar type valley glaciers in the Expedition Area corroborate the above statements. Of course, subglacial channels are more or less absent because of the negative ice temperatures, but marginal erosional features are present in remarkable variety.

a) Marginal Channels

As Sissons (1961, p. 15) and Ives and Kirby (1964, p. 920) rightly emphasize, it is premature at the present state of research to put up a new general system of glacial drainage channels. A special classification of the observed erosional features is therefore not attempted here. Only the terms used are defined. "Marginal" (German: "randlich") is used with similar meaning to that in the classical papers on Alaskan glaciers by Tarr (1908, p. 100, 1909, p. 141-142) and von Engel (1911, p. 109ff) and as in the more recent works of Charlesworth (1957, p. 458, 460), Sissons (1958a, p. 161), Derbyshire (1962, p. 1124), Price (1960, p. 486). It is about synonymous with "strictly lateral" as defined by Mannerfelt (1945, p. 17, 228; 1949, p. 197; 1960, p. 109) and similar to "lateral" as used by Schytt (1956, p. 67ff although the genesis might be different here), Henderson (1959, p. 25, 43), Ives (1959, p. 58), Hoppe (1957, p. 15-16), Ives and Kirby (1964, p. 920), Syrlä (1965, p. 22) and others; but the "lateral" of Rich (1908, p. 528) is quite different. "Border drainage channels" (Logan, 1945, p. 12), "glacier-margin channels", "ice-margin channels", "ice-marginal channels" (Flint, 1957, p. 161, 166) seem to denote similar features as the ones described here.

In the Expedition area, marginal rivers were observed to flow between ice and valley wall (being the most common case); in channels cut into the valley slope but following closely the ice edge; between ice and lateral moraine; and between lateral moraine and valley wall.

Figures 68-70, 74-76 and 78 show the varying results of the erosional activity of the same marginal river in a relatively short stretch of some 800 m along the Crusoe Glacier. The main factors responsible for these differences are here local variations in the lithology, in the steepness of the valley side and in the grade of the river bed. In Figures 68 and 69 two "double-walled" marginal channels (Tarr, 1908, p. 100) are shown, one is rather shallow and fairly broad and the other is deep and narrow. Both were cut by the Accident River into the gently sloping valley side only some 150 m apart, but as the slope material and the grade of the river bed were not the same the channels developed differently. In Figure 70 the valley slope bordering the ice margin is steeper than in the previous examples. The river had cut a narrow trench between the ice and the valley wall by mech-

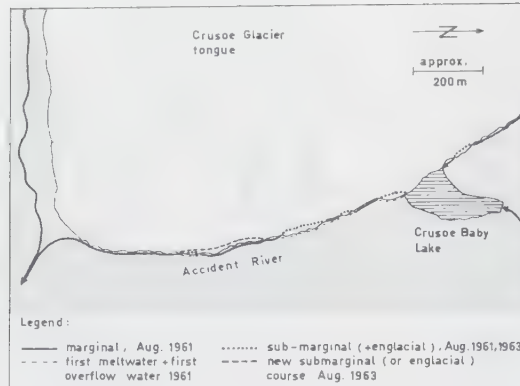


Fig. 72: Marginal and sub-marginal stretches of Accident River.

Fig. 73: Development of marginal and sub-marginal channels at a steep valley side (diagrammatic cycle).

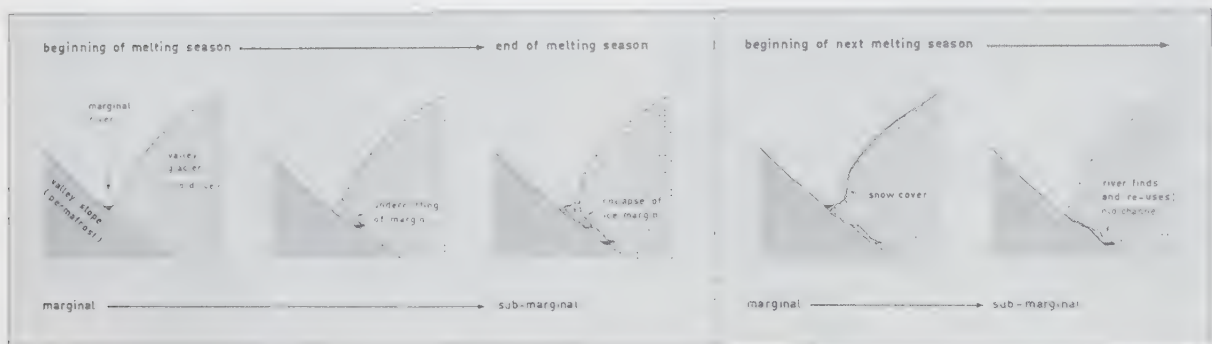


Fig. 74: Undestroyed marginal channel stretch at the left margin of Cruso Glacier cut in previous summers by Accident River. Rod (3.50 m) for scale (23 June 1963).

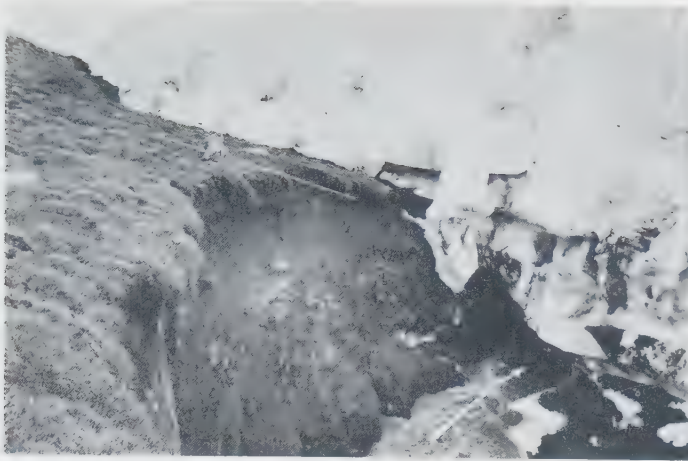


Fig. 75: Left margin of Crusoé Glacier. Marginal river (arrow) becoming sub-marginal. Rod (3.60 m) and rucksack for scale (16 July 1963, 2300).



Fig. 76: Left margin of Crusoé Glacier. Accident River emerges from englacial (1) and sub-marginal (2) passages to continue in marginal channel (3). Rod (3.20 m) for scale (15 July 1963, 2100).

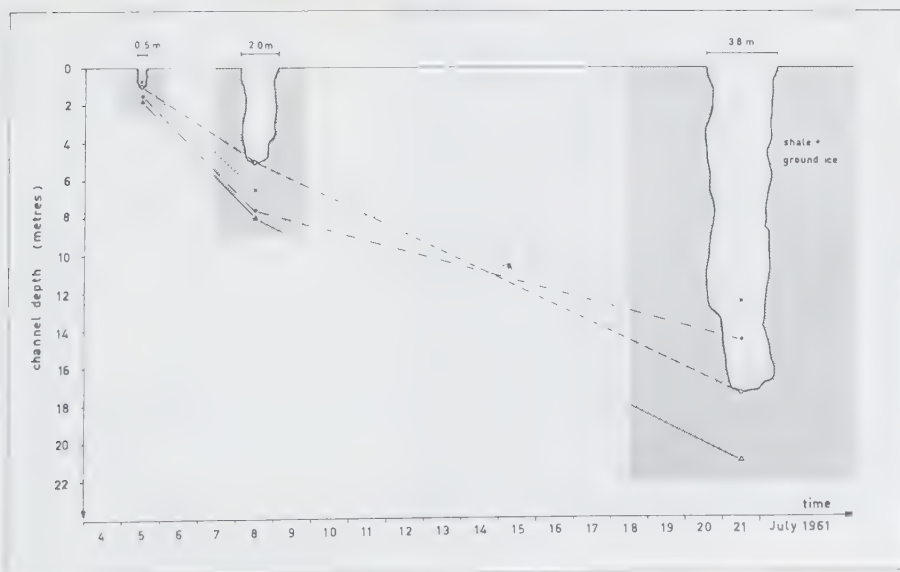


Fig. 77: Progress of channel erosion (bed of Accident River) recorded at 4 different sites.

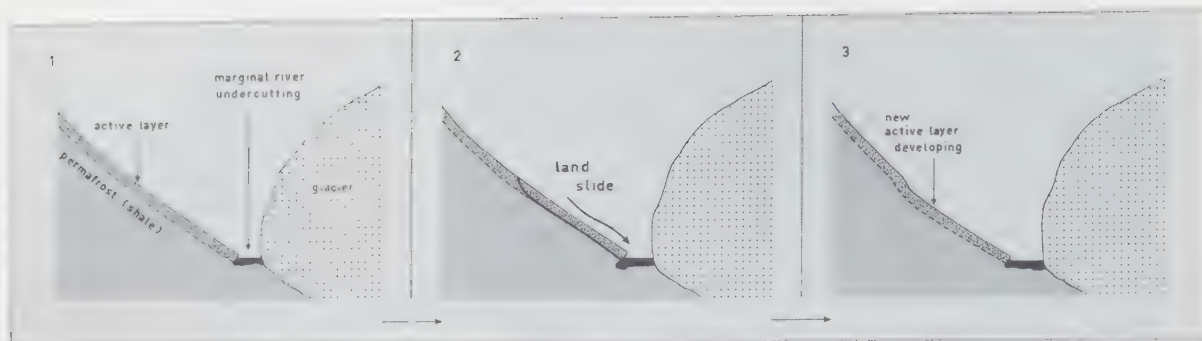


Fig. 78a: Erosion of permafrost slope by undercutting by marginal river (diagrammatic).

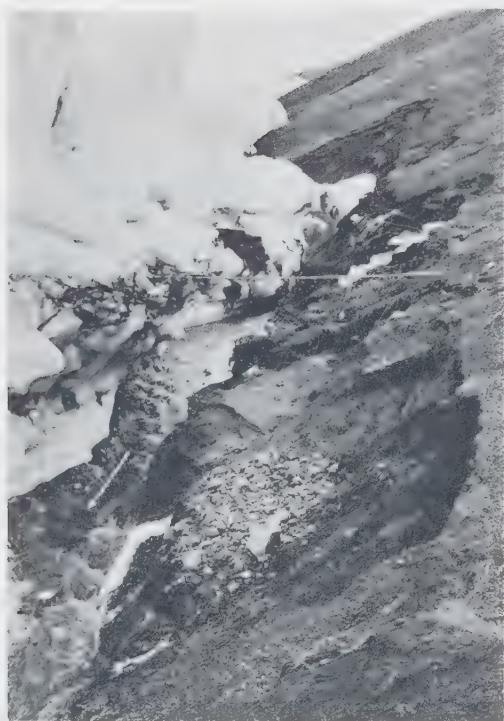


Fig. 78b: Slumping of active layer of permafrost slope after undercutting by Accident River (arrow) at left margin of Crusoe Glacier. Rod (5.50 m) for scale (8 July 1961).

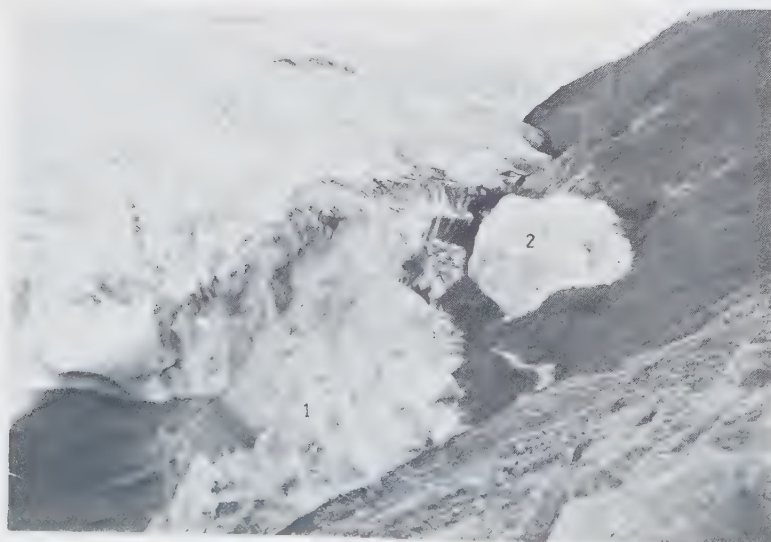


Fig. 79: Right margin of White Glacier with partly collapsed marginal channel (1) and cut-off ice piece (2) (14 July 1963).

anical erosion and thawing. Channels with similar cross-sections have been observed at various sites along the Crusoe, Thompson and White glaciers (Fig. 23). Such "one-sided" or "single-walled marginal channels" (Tarr, 1908, p. 100) also occurred at places with a more gently sloping valley side.

b) Sub-marginal Channels

"Sub-marginal" is here taken to mean beneath the margin of a glacier but close, and more or less parallel, to the ice edge. The term has been used in a similar way (but not always with this precise meaning) by von Engel (1911, p. 110), Sissons (1961, p. 15), and Derbyshire (1962, p. 1124), and is roughly synonymous with "sub-lateral" defined by Mannerfelt (1945, p. 18, 228; 1949, p. 197; 1960, p. 110), and is used by Ives (1959, p. 58) and others; whereas Rich's (1908, p. 528) and Charlesworth's (1957, p. 460) usage of "sub-marginal" is completely different. Stone's "subglacial marginal channels" (1963a, p. 31-32; 1963b, p. 340-341) are not fully comparable although they show some similarities.

Figures 23, 24, 42, and 44 show that on 9 July 1961 the glacier margin that had been deeply undercut by the Accident River collapsed, burying the river course. In each of the years 1962, 1963, 1964 and 1965 a similar collapse of the marginal channel took place. Such icefalls, caused by the excessive undercutting of flowing water, were observed not only at this lake outlet channel but also at various places along the glacier margin, even at sites where the valley walls were less steep. The result was often a temporary damming up of the river in a small marginal lake. Usually, as was the case at the site shown in Figure 24, the dammed-up water sooner or later found a way through the icefall and the river continued to flow sub-marginally. If the ice blockage was rather tight the impounded river would by-pass it by overflowing the bank. The erosion terraces in Figure 71 are the result of such an overflow.

It is known from present temperate glaciers in the Arctic and Sub-Arctic and has also been suggested in various studies on fossil channels, that strictly marginal channels are rare and mostly not continuous, that meltwater rivers tend to cut under the margin of the glacier, especially if the valley slope is steep, and that they continue sub-marginally or subglacially; it has also been said that sub-marginal or even subglacial drainage is the normal case (Tarr, 1908, p. 100; von Engel, 1911, p. 109; Mannerfelt, 1945, p. 17-23, 223; 1949, p. 194, 197; 1960, p. 110; Sharp, 1947, p. 50; Sissons, 1958a, p. 163; 1960, p. 131, 138-139; 1961, p. 15-16; Ives, 1959, p. 58, 60, 62; Henderson, 1959, p. 43; Price, 1963, p. 137; Syrlå, 1965, p. 22-24).

On sub-polar type glaciers also, despite the cold ice, the undercutting of the glacier edge by marginal rivers has been observed (Maag, 1963, p. 153, 158; Ives and Kirby, 1964, p. 920). In Figures 72 and 88 the marginal and sub-marginal (or englacial) stretches of two glacial rivers are shown. Similar situations were also observed along the margins of the Thompson Glacier and other glaciers on Axel Heiberg Island. The rivers along these sub-polar type valley glaciers flow sub-marginally at several places, nevertheless the extent of strictly marginal drainage is still greater than with temperate glaciers. Even when the rivers flow below the ice margin they probably follow it more closely. In general, true subglacial drainage does not seem to occur.

Figure 72 shows that at the beginning of the melting season, the water flowed for most of its course strictly along the margin of the glacier, whereas later in the summer at various places the river changed to a sub-marginal course, either because of the undercutting and subsequent collapse of the ice-margin or because it re-discovered old sub-marginal or englacial tunnel stretches which had not been destroyed by the pushing of the glacier. This was also observed on other glaciers in the Expedition Area. Figure 73 depicts diagrammatically such a development. The observation that in early summer more rivers flow strictly marginally than at the end of the summer more or less agrees with the findings of Tarr (1909, p. 103) in Alaska and with suggestions of Mannerfelt (1945, p. 17, 74, 102; 1949, p. 196) which were accepted by Sissons (1958a, p. 164), Ives (1959, p. 58) and Derbyshire (1962, p. 1125), then later questioned by Sissons (1961, p. 19). The contrary development has also been observed on occasion, i.e. short sub-marginal channel stretches cut or re-used early in the summer were blocked later in the season when the river had again to follow a marginal course or the roof of the tunnel collapsed and the river continued to flow in an open channel.

Sections of a channel may stay open and be re-used by the meltwaters of subsequent years.



Fig. 80: Right margin of White Glacier with edge-rounded boulders in bed of former marginal river; fine material washed out. (22 August 1963).



Fig. 81: Meltwater cuts (1, 2, 3) in crest of White Glacier terminal moraine; concentration of coarser material due to washing out of finer fractions. (26 August 1963).



Fig. 82: Edge-rounded boulder (3 m high) with morainic material on top; terminal ablation moraine of White Glacier. Supraglacial river in background removed the fine material from around boulder, (21 June 1963).



Fig. 83: Ice-mound on White Glacier tongue covered by well-sorted and rounded pebbles deposited by a supraglacial river, (21 June 1963).



Fig. 84: Esker-like, well-rounded deposits develop on top of terminal moraine of White Glacier. For scale rucksack on top of mound, (21 June 1963).



Fig. 85: Snout of White Glacier and terminal moraine (with meltwater ponds and erosional cuts) and outwash plain; large gap in foreground has been cut mainly by the waters from the emptying Between Lake. (29 July 1963, 1700).



Fig. 86: Scars in lateral moraine of White Glacier caused by Ermine River. (17 August 1961).

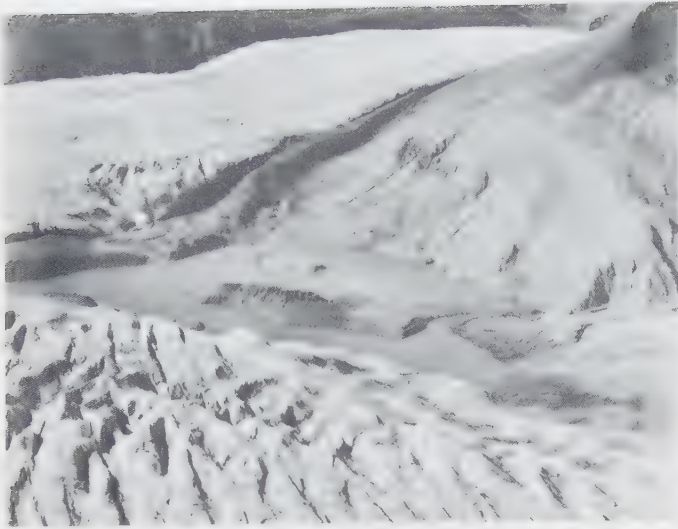


Fig. 87: Between Lake ponded by White Glacier (background) and Thompson Glacier (foreground). Shorelines and stranded ice pieces indicate highest water level. Gap in lateral moraine ridge eroded by river leaving White Glacier margin (oblique air photo, July 1962).

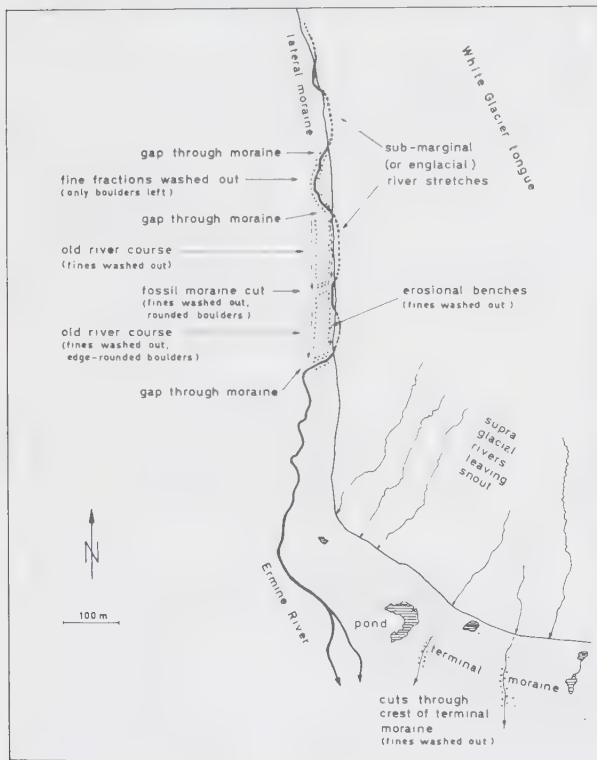
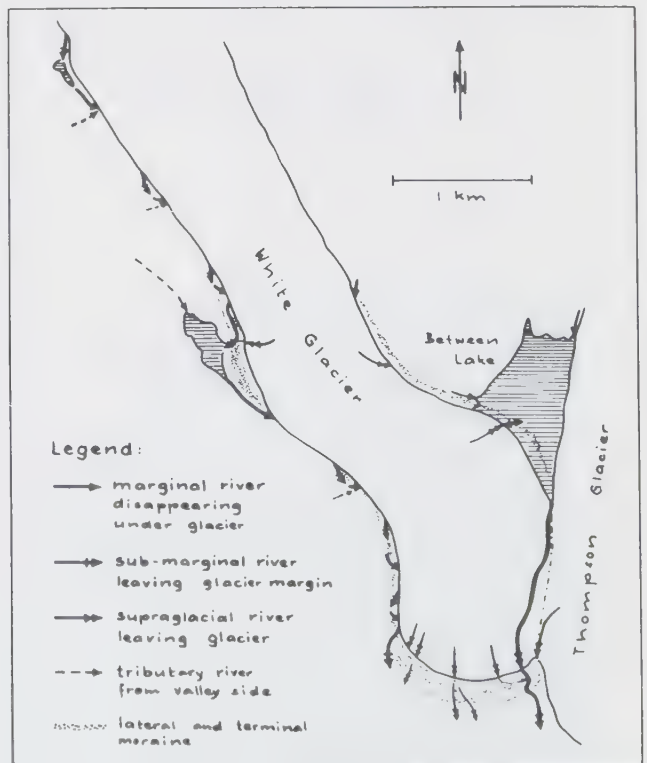


Fig. 89: Meltwater rivers flowing off and disappearing under margin of White Glacier, summer 1963.

Fig. 88: Cuts through White Glacier moraine by marginal and supraglacial rivers, summer 1963.



This was seen at Crusoe Glacier where in June 1963, before the meltwater started to run, several stretches of the 1961 and 1962 marginal channel of Accident River were found undestroyed (Fig. 74); later the river re-occupied them. Probably sub-marginal channels can also stay open although this will occur less frequently.

It has been suggested that stagnant and shrinking glaciers would show more extensive systems of sub-marginal or englacial run-off channels than actively moving and growing glaciers where marginal drainage would dominate (von Engel, 1911, p. 109; Mannerfelt, 1956, p. 15-16, 223 and references therein; Sissons, 1961, p. 18-19; Clayton, 1964, p. 107-112; Vivian, 1965, p. 393). The situation at some stretches along the White Glacier confirms this to a certain extent (Chapters B and C). On the other hand, along the actively sideward-pushing and advancing Thompson and Crusoe glaciers, relatively long stretches of sub-marginal passages were found (Figs 72, 75, 76), but then again along the thinning White Glacier the river flowed strictly marginally at several places (Figs 88, 89).

There are differences from year to year in the distribution of sub-marginal and marginal stretches of the river course along the Crusoe Glacier (Fig. 72) probably mainly because of slight variations in the dynamics of the glacier and consequent changes in the closing of certain channel stretches. But, on the whole, the sites where the water enters or leaves the glacier edge are nevertheless remarkably stable over the years (Chapters C and D). Okko (1956, p. 81) has somewhat similar observations from Iceland. Some of the sites where glacial rivers were observed to enter or leave the ice margin seemed to coincide with: structural lines, as at "Ice Cave" (Chapter B); abrupt lithological changes; abrupt steepening of the valley wall, spurs; places where ravines joined the glacier from the side, some of them containing small, steep "land rivers" with relatively warm water which might have helped in the undercutting of the ice margin (von Engel, 1911, p. 111, 124, 134; Mannerfelt, 1945, p. 23; Liestøl, 1956, p. 145-146; Sissons, 1961, p. 20; Gjessing, 1966, p. 141, 143); and places where a river, flowing between valley slope and lateral moraine, cuts through the moraine towards the glacier margin (Chapter C). It is interesting to note that the ice margins frequently collapsed at approximately the same places each year.

c) Erosive Power of Marginal and Sub-marginal Rivers

Various authors mention the great erosive power of marginal glacial rivers arising from their steep grade, great volume, frequent shortlived flood discharge, turbulent water, heavy load, etc. (Brunhes, 1906/07; Tarr, 1908, p. 100; von Engel, 1911, p. 120-129; 1945, p. 103; Mannerfelt, 1945, p. 49; 1960, p. 112-113; Sharp, 1947, p. 50; Flint, 1957, p. 166, 168; Charlesworth, 1957, p. 461; Rudberg, 1963b, p. 142-3; 1965, p. 4, 16), but it has been suggested that the rates of cutting vary considerably (Flint, 1957, p. 168). These statements are corroborated by the observations in the Expedition Area.

On 7 July 1961 two water samples were taken from the marginal Accident River, one just below the outlet of Crusoe-Baby Lake and the other some 1000 m further downstream at a point near the snout of the glacier. The suspended matter in the river water was found to have increased over this relatively short distance from 21 mg to 3727 mg and the total amount of dissolved solids from 41 mg to 106 mg per 1000 cm³ of water. This is an indirect indication of effective erosion of the valley slope by the marginal river, especially as the petrographic composition of the suspended material suggests that it must have been picked up mainly from local outcrops along the marginal channel. Figure 77 shows how fast a short stretch of a double-walled marginal channel (Fig. 69), cuts through a broad spur, was deepened and enlarged by the waters of the Accident River in July 1961. The erosive work is remarkable, even taking into consideration the steep grade of the river, the fact that the bedrock was friable though frozen, that there was shale with thick layers of ground-ice, and that the Accident River was first dammed up behind the spur and therefore probably heated up a little increasing its melting capacity. However, only some 200 m downstream, where the grade of the river is much less and where frozen morainic and gravelly material replaces the shale and ground-ice, the erosion is much less spectacular (Fig. 68).

It has been suggested that where a glacier lies against a steep slope erosion of the valley side would be slight, or even absent, because the marginal river would first erode the ice (Visser, 1928, p. 189, in Mannerfelt, 1945, p. 19; Holdar, 1959, p. 234; Sissons, 1961, p. 16). A comparison of the channel cross-sections of 5 and 7 July 1961 in Figure 23 shows that during these two days it was, of course, mainly the ice that was melted by the marginally outflowing lake water;

nevertheless, a substantial part of the slope was taken away. It must be noted, however, that the material was not very resistant. This was observed in a marginal stretch of the channel, but could also occur in sub-marginal ones where it might be even more vigorous because of possible siphon effects. At several places along the Crusoe and Thompson glaciers it was observed that the marginal rivers flowing in single-walled channels were undercutting the steep mountain side and were stripping the active layer off the permafrost, occasionally causing landslides and mudflows (Fig. 78). The effect of erosion along a steep valley side was further evidenced by a marked break in the gradient of the mountain slope at and below the ice edge, e.g. on the right margin of the White Glacier, some 200 m downstream from Ice Cave. This probably resulted from the action of marginal and, more especially, sub-marginal meltwaters which eroded each summer for years at approximately the same level.

Below the Ice Cave, where the marginal ice surface of the White Glacier is relatively flat, it was observed in the summer of 1963 that the Ermine River flowed for a short stretch over the ice and had cut through the ice down to the slope beneath (Fig. 79); this superimposition seems to be a similar process to that suggested by Price (1960, p. 486-88; 1963, p. 136, 138; 1964, p. 11).

As most channel stretches are used not only during one summer, but re-occupied in subsequent years, the repeated work of the marginal and sub-marginal rivers finally has some erosional effect even on steep valley slopes. This produces "washed hillsides" (Charlesworth, 1957, p. 460) and washed moraines (Chapter C 2), and "benches" (Charlesworth, 1957, p. 460; Sissons, 1958a, p. 161; Price, 1960, p. 486; Derbyshire, 1962, p. 1124) which are also called "terraces" (Mannerfelt, 1945, p. 228; Sharp, 1947, p. 50). They are illustrated in Figures 142 to 144.

Further examples of the action of meltwaters at the margin and in the tongue area of valley glaciers, especially during floods, are discussed in section 3 below.

Many of the channels considered in this study were cut directly by outflowing lake water, or their development was to some degree influenced by the drainage of ice-dammed lakes. This could lead to the false conclusion that ice-dammed lakes are necessary for the formation of marginal and sub-marginal channels, as was assumed by many authors after Kendall's paper of 1902 until fairly recently (Drehwald, 1955). It must, however, be emphasized that there is plenty of evidence in the Expedition Area to show that the normal run-off from the glaciers and the water from the surrounding valley sides suffice to erode these channels (Mannerfelt, 1945, p. 224; Peel, 1956, p. 486; Sissons, 1958a, p. 159, 177-85; Ives, 1960, p. 60; Price, 1960, p. 483-86; 1963, p. 133, 136; Derbyshire, 1962, p. 1122, 1125). Of course the water from emptying ice-dammed lakes is also very important for the formation and modification of drainage channels, as demonstrated in this study.

Evidence of the depositional activity of marginal and sub-marginal rivers - although present at many places in the Expedition Area - is not discussed here, except for some remarks on lake deposits and reworked moraines.

2) Influence of Meltwaters on the Moraines of Arctic Valley Glaciers

Some occasional observations on the effect of supraglacial and marginal meltwater on present-day moraines of sub-polar type valley glaciers were made in the Expedition Area. The results of the work of the water can vary very much, ranging from moderate modifications of morainic material such as washing, sorting, rounding and re-depositing to incisions through moraine ridges and complete destruction of whole sections of moraines.

Modifications to moraines by meltwater within the ablation zone of the glacier, and not only in the proglacial area, and the resulting complexity and variety of deposits and forms have been mentioned by various authors referred to below, but most of their statements are based on observations from temperate glaciers or on investigations of fossil moraines. The examples presented here show that also on and along present-day sub-polar type valley glaciers the meltwaters help to determine the structure and shape of the moraines.

a) Modifications of Morainic Material by Meltwater

Figures 80 and 81 show a concentration of coarse, partly edge-rounded pebbles and

boulders in the channel at the foot of the right lateral moraine of the White Glacier and in a cut through the crest of the terminal moraine. The marginal, supraglacial and proglacial rivers have washed out the finer fractions in these places, whereas the larger components of the moraines were partly blunted at the edges but otherwise left undisturbed. Similar observations have been cited by von Engel (1911, p. 134), Ahlmann (1933, p. 177, 180), Pawlowski (1938, p. 239), Logan (1945, p. 12), Holmes (1949, p. 1432), Sharp (1949, p. 289, 294, 313), Woldstedt (1954, p. 95), Portmann (1954-55, p. 17; 1956, p. 425; 1964, p. 131), Okko (1956, p. 8, 36, 37, 55, 97, 110), Lister and Wyllie (1957, p. 48), Flint (1957, p. 75, 109, 121, 122), Jayet (1958, p. 345), Henderson (1959, p. 19-20), Schytt (1959, p. 221), Kirby (1961, p. 15-16), German (1962, p. 124, 134), Tricart and Cailleux (1962, p. 347, 361, 362), Weidick (1963, p. 28, 37), Fulton (1965, p. 568).

The large boulder in Figure 82 would appear to have been once covered by finer morainic material as witnessed by sand and small stones on its top. The block was too large to be moved by the supraglacial meltwater rivers running down the tongue of the White Glacier but the smaller material was removed and the edges of the boulder were rounded.

The highest proportion of well-rounded and sorted glacial debris was observed near and at the tongue (Sharp, 1949, p. 295). Such water-worn pebbles that have been transported and deposited by a supraglacial river and that now cover an ice-mound near the snout of the White Glacier are shown in Figure 83. As the glacier is melting back at its tongue, eskerlike, well-rounded and sorted material is deposited on top of the terminal ablation moraine (Fig. 84). The forms that develop are similar to the ones described by John and Sugden (1962, p. 359) from Iceland. Some of the material found in the upper part of the ablation area showed a grain-size distribution and a roundness comparable to those of glacio-fluvial material collected in the proglacial Expedition River outwash plain. Rounding and sorting of glacial debris has been described from many temperate glaciers and fossil moraines, but most of these observations concern supraglacial and terminal debris, whereas on Axel Heiberg Island marginal deposits are also greatly modified by meltwater (Ahlmann, 1933, p. 177; Pawlowski, 1938, p. 239; Logan, 1945, p. 12; Sharp, 1947, p. 34, 36, 46; 1949, p. 293-95, 299, 313; Cailleux, 1954, p. 305; Portmann, 1954-55, p. 17; Okko, 1956, p. 38, 55, 59, 64, 97; Charlesworth, 1957, p. 406, 408; Flint, 1957, p. 115, 138; Tricart and Cailleux, 1962, p. 343, 345, 346, 362; Ruchina, 1958, p. 587; German, 1962, p. 134-35).

The meltwater rivers were observed to re-deposit some of the fine-grained material they had eroded up-glacier into shallow ponds situated on the terminal moraine of the White Glacier (Fig. 85), on the ice-surface, or along the lateral moraines. This process with its resulting pockets of stratified sediments was recognized some 125 years ago on alpine glaciers by Agassiz (1841, p. 201, 203).

Part of the left lateral moraine of the White Glacier is completely covered every year by the rising waters of Between Lake, it emerges when the lake drains (Fig. 13). The differences between the temporarily drowned and the non-submerged parts of the moraine were clearly visible on the ground and are detectable also in the aerial photographs (Figs 87, 140). A cover of fine-grained stratified lake sediments of varying thickness was deposited (Chapter B). Therefore these moraines showed higher percentages of fine material than "normal" ones, especially in the lower parts. This confirms observations in Greenland by Lister and Wyllie (1957, p. 47). On the other hand, the action of waves, currents, and the fast changing lake level caused some local washing out of the finer grain fractions.

As the conditions controlling the deposition of material on, along, and at the end of a valley glacier are different from place to place and constantly changing (Tarr, 1908, p. 99; Okko, 1956, p. 38; Ives and Andrews, 1963, p. 29) sediments of remarkable variety and complexity accumulate. A good example of this is the terminal moraine of the White Glacier. Various stretches of the lateral moraine show a similar diversity of sediments, grading from heterogeneous, non-sorted, angular glacial debris to stratified, well-sorted and well-rounded "glacio-fluvial" deposits. Other ice-marginal accumulations, such as kame deposits, "Moränenterrassen" (Eggers, 1961), etc., are further evidence of the activity of rivers from the glacier and the valley side, and add to the complexity of the sediments found along the White Glacier and other glaciers in the same area; however, they are not discussed here.

b) Examples of Erosional Effects

On and beside the White Glacier whole series of erosional effects were observed:

- removal of the smaller particle fractions, leaving the larger ones more or less in place (Figs 80-82),
- small erosional terraces (benches) cut along the lateral moraine ridges by marginal streams flowing between them and the ice (Fig. 140),
- undercuts of the frozen, ice-cored lateral moraine ridges, mainly formed by thermal erosion of the marginal river,
- scars caused by the down-sliding of surface material, especially where marginal meandering meltwater had undercut steep moraine slopes (Fig. 86),
- cuts across the crests and incisions through the entire ridges of lateral and terminal moraines (cf. observations from Spitsberge, Iceland, Alaska/Yukon, e.g. Pawlowski, 1938, p. 240; Gripp, 1938, p. 224; Sharp, 1947, p. 46; Okko, 1956, p. 64, 68, 110; Todtmann, 1960, p. 43, 61, 62, 77, 78; Szupryczynski, 1963, p. 37, 48, 105, photographs and figure),
- stretches where moraine ridges are missing, due to removal by floodwaters, or to rivers and lakes hindering the accumulation of new morainic material.

Erosional cuts across the lateral and terminal moraines were often seen to occur where (Figs 82, 85, 87-90): a) supraglacial rivers reached the ice margin and joined the marginal or proglacial rivers, b) tributary rivers from the valley side reached the moraine, c) a sub-marginal river left the ice margin, d) a marginal river running between valley wall and moraine ridge curved back towards the ice margin often becoming sub-marginal at this point, or e) a river was dammed up by icefalls and spilled over the moraine crest.

It has already been mentioned that, despite minor variations in the régime of the rivers and the glacier, the places where marginal rivers enter or leave the ice margin do not appear to change much over the years. Also, supraglacial meltwaters were observed to use roughly the same beds year after year (Holmes, 1955, p. 22-23) and to leave the glacier at the same points (Chapter D). The repeated erosional impact may finally result in cuts through the crest or the whole of a lateral or terminal moraine.

In Figures 81, 85, 89 and 90 some of the cuts across the terminal moraine of the White Glacier are visible. The study of aerial photographs of other valley glaciers on Axel Heiberg Island with similarly well developed terminal moraine ridges suggests that such a radial pattern of cuts seems to be fairly typical for this group of sub-polar type glaciers. As the major part of the meltwater and of the draining lake water runs off marginally, the gaps at both ends of the moraine arch, where it joins the lateral moraine ridges, are sometimes larger than the cuts across the middle part of the terminal moraine because the latter is eroded by the relatively smaller supraglacial rivers that flow down the tongue and may often only be powerful enough to cut into the moraine crest. Hence, the whole middle section of the moraine arch of these cold valley glaciers, although displaying some minor cuts, remains more or less intact. Robitaille and Greffard (1962, p. 93) reached similar conclusions regarding the preservation of the Thompson Glacier push-moraine. In contrast, the terminal moraine of many valley glaciers in temperate regions is often most eroded in the middle section as the major part of the run-off is sub- and englacial and leaves the tongue from a more or less central glacier door, although additional meltwater channels of minor importance may exist (Lütschg-Loetscher, 1950, p. 17-18; German, 1962, p. 123-24, 140).

In several papers on glaciers in the Sub-arctic and the Arctic, floods, such as those produced by the outbreak of an ice-dammed lake, are considered to be responsible for the cutting of gaps through moraines and for the destruction of whole stretches of glacier margin deposits, as well as for stripping off the moraine cover over wide areas and for the creation of "meltwater trim lines" (Thorarinsson, 1939b, p. 229; Okko, 1956, p. 68; Lister and Wyllie, 1957, p. 45; Todtmann, 1960, p. 43, 61, 62, 64; Szupryczynski, 1963, p. 48, Fig. 8; Ives and Andrews, 1963, p. 32-33; Price, 1964, p. 10, 11, 22). In section C 3 it is shown that the water from emptying ice-dammed lakes can indeed have large erosional effects. However, probably most of the gaps in the moraines of the Expedition Area are the result of the normal, annually repeated, activity of the rivers; though, occasional floods may have enlarged these cuts or initiated new ones.

It has been said that the presence of marginal lakes and the activity of meltwater rivers is influential, especially along steep slopes and at the glacier terminus, in preventing the formation of lateral and terminal moraine ridges (Bretz, 1935, p. 179, 182; Sharp, 1947, p. 49, 50; Flint, 1957, p. 135; Adams, 1966, p. 1). In addition, rivers and rivulets running down the valley wall assist in the modification and destruction of the lateral moraines (Sharp, 1947, p. 49). The fore-

going may be an explanation of the scarcity of lateral and terminal moraines of the advancing Crusoe Glacier. This is, of course, just one possible answer to the complex problem as to why some valley glaciers on Axel Heiberg Island have well developed lateral and terminal moraines, whereas others have none or only poor ones.

The effect of the meltwater is especially large on, along, and near the terminus of inactive, slowly moving or stagnating glaciers which are thinning and receding or even decaying (Ahlmann, 1933, p. 177; Sharp, 1949, p. 1949, p. 293; Flint, 1957, p. 160, Tricart and Cailleux, 1962, p. 343, 418). On Axel Heiberg Island, a good example of this is the Iceberg Glacier (approximately $79^{\circ} 30' N$, $92^{\circ} W$), a large outlet glacier some 30 km east of the Expedition Area, reaching with its snout the tidewaters of the fiord. It shows signs of intensive meltwater action on the slightly inclined surface of the tongue as well as in the marginal deposits (Fig. 91). White Glacier, from where most of the observations are reported in this section, is also thinning and receding, but it is certainly not a passive glacier (Müller, 1963g, p. 68-73). Also on and along actively moving and advancing glaciers, such as the Crusoe and Thompson glaciers, meltwaters work with great effect; their influence on the moraines should not be underestimated.

The total meltwater production of a sub-polar type valley glacier, such as the White Glacier, is normally smaller than that of a comparable glacier in more temperate regions. But this must not necessarily imply, as statements of Flint (1957, p. 160) and Okko (1956, p. 119) seem to do, that the glacial debris on and along such a glacier becomes less modified by meltwater. As practically all the water runs off supraglacially and marginally, during the ablation season more meltwater is probably concentrated on the surface and along the glacier margin than on and along a comparable temperate glacier, e.g. in the Swiss Alps, where most of the water soon takes an end-and/or subglacial course. It is therefore suggested that on - and especially along - a sub-polar type valley glacier there are more signs of water action, such as modified morainic material, erosional terraces and cuts than on and along a temperate glacier.

3) The Effects of Flood-water in the Marginal and Proglacial Area

In the marginal and proglacial river beds the effectiveness of high discharges resulting from rain storms or the draining of ice-dammed lakes was investigated. Such floods, though mostly of only moderate proportions, recur with sufficient frequency to be of great geomorphological importance (Leopold, Wolman and Miller, 1964, p. 66).

a) Methods of Investigation

To obtain quantitative data on the changes in the shape of the river beds, some 13 profiles across channels and river beds and across part of the outwash plain were surveyed two or more times during each of the summers of 1961 and 1963. In the case of the small channels a horizontal levelling rod was laid across the bed from shore to shore, and depth measurements were taken at 10 or 20 cm intervals to give the profile. The wider profiles were levelled with a Wild theodolite. The standard errors in the relative positions of the levelled points are estimated to lie within the drawing accuracy. As it was generally not possible to enter the river bed during high discharge, but only before and after, progressive changes could not be recorded but only the net-change. Series of photographs, taken from marked standpoints, and the comparison of air photographs from different summers, give valuable supplementary evidence of the modifications in the river beds and river courses.

The accuracy of the discharge data varies very much. For the Ermine River a good discharge record exists for the first part of the 1961 summer. W.P. Adams used a self-built dam with a weir and an automatic level recorder to gauge this river, checking with a current meter (Adams, 1966, p. 5, 7-8). Unfortunately, the installations were destroyed by a flood discharge during a rain storm in July 1961; for the remainder of the 1961 summer and during 1963, the discharge was estimated, except for a few measurements with the Sodium Dichromate Dilution Method (Adams, 1962, p. 61-55; 1963, p. 53). The installation of a similar gauging station in the Between River was not possible because of its considerable larger size, the special nature of its banks and the lack of a stable cross-section. The discharge had then to be estimated. Surface velocities were measured with floats. The mean velocity in the vertical was taken to be approximately 0.85 of the float velocity (Arnborg, 1955a, p. 188; Rudolph, 1962, p. 132; Fahnestock, 1963, p. A30; Ostrem, 1964, p. 22). The discharge figures thus obtained are very crude because of the ever changing velocity

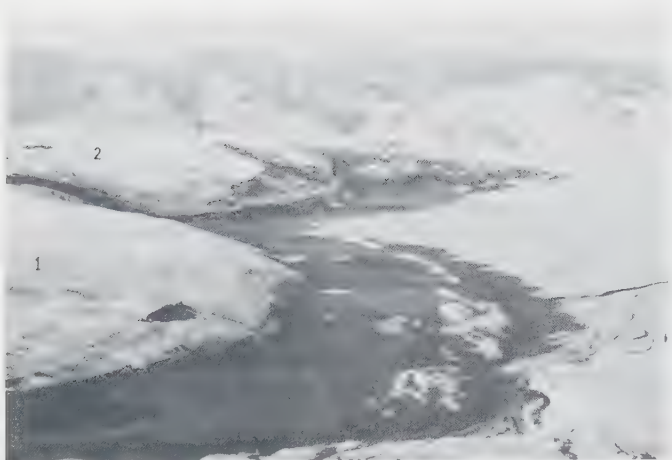


Fig. 90: Snouts of White (1) and Thompson (2) glaciers (foreground); Ermine River marginal to White Glacier, push-moraine in front of Thompson Glacier. Cuts in terminal moraine of White Glacier. (21 July 1963, 0100).



Fig. 91: Iceberg Glacier (approximately $79^{\circ} 25'N$, $92^{\circ}W$), some 5 m wide. Meltwater action on ice and in marginal area of thinning and receding glacier (vertical air photo, RCAF A 16864-76, 13 August 1959).

Figs 92a & b: Gauging site (x) in lower Ermine River bed:
a) during first flood, 14 July 1961, 0600, dam with weir and level recorder.
b) after floods, 19 August 1961, weir and level recorder destroyed and carried downstream.



distribution across and along the river (Rudolph, 1962, p. 134, 142-143, 169; Leopold et al., 1964, p. 166), and because of the instability of the cross-section which could only be surveyed during low stage and not during flood periods when most changes occur. Nevertheless, these estimates are of some use, and at least for the low stage they could be checked against two measurements carried out by W.P. Adams with the dilution method; the results compared favourably. At high stage, the discharge of the Between River was fairly well known as more than 90% of the water originated from Between Lake whose drainage volume was measured (Chapter B). For future research in turbulent, muddy rivers like the Between River, where the installation of an automatic level recorder is not possible, the most suitable way of determining the discharge would be by electrical conductivity measurements as described by Ostrem (1964).

Some 70 water samples were collected from different glacial rivers. Most of them were sampled simply by holding a polyethylene bottle of 500 or 1000 cm³ capacity obliquely in the river and letting it fill slowly. Hjulström (1935, p. 388) considers this improvised method gives fairly accurate results. But only occasional sampling was possible and in most instances only surface water near the shore could be reached safely. The value of such isolated samples has been questioned (Gogarten, 1909, p. 273; Hjulström, 1939, p. 68; Sundborg, 1956, p. 233, 293; Rudolph, 1962, p. 199-200). It is known that the concentration of suspended matter, as well as its grain size distribution, varies considerably in time and within the cross-section, and also with the degree of turbulence. On the other hand, in fast flowing, highly turbulent water - and this was usually the type in the Axel Heiberg rivers - the sediment, especially the finer fractions, is distributed more equally. Nevertheless, one has to be very cautious when comparing and interpreting the various samples. However, as data on the suspended and dissolved load of glacial rivers in the Arctic are scarce, some of the results are presented here.

Different methods were used for separating the suspended load from the water. About half of the samples were filtered in the field through a Munktell filter paper (class I, No. OA). The rest were taken to Zürich, where the suspended sediment was separated by decantation and centrifuging. The filtered samples are not truly comparable with the other samples, as the grain sizes and their amounts lost through the filter are unfortunately only approximately known. After separation, the suspended material was dried in an oven at 105°C, and the total weight was determined. Grain-size distributions were analysed with a Sartorius sedimentation balance at the Mineralogical-Petrographical Institute of the University of Berne, and with a microsedimentation-analysis apparatus at the Swiss Federal Institute for Testing Materials and Research (EMPA). Discussion of the limitations of these methods and of the margin of error is given in Niklaus (1967, p. 74-75), and Gessner (1963, p. 133-136). Although flocculation was no problem, some "Calgon" (sodiumhexametaphosphate) or sodiumpyrophosphate was added to guarantee complete dispersion.

The determination of the total amounts of dissolved substances (dry residue) and some analyses of the chemical constituents was done at the Kantonales Laboratorium Zürich according to the standard methods in use there (Thomas, 1948, p. 97-98). In most cases the suspended material and the water with dissolved substances could only be separated several months after the sampling had taken place. Because of this, the total amount of dissolved solids and the relative concentration of the different constituents might have changed a little; such an error is not considered to be serious.

No measurement was made of the amount and size of the bed load in the Ermine River or the Between River during flood periods, as this would have been too difficult and too dangerous an undertaking (Fahnestock, 1963, p. A26-27; Hubbell, 1964; Leopold et al., 1964, p. 185; Frécaut, 1966, p. 327; King, 1966, p. 87). But an attempt was made to estimate at least the net volume change in the Ermine and Between river beds by comparing the cross-sections surveyed before and after the flood periods and by consulting various photographs. In addition, to get an idea of the size of the largest boulders moved by the flood waters a project was undertaken together with S. Rudberg (1963b, p. 147-148) and W.P. Adams (1966, p. 68). At the beginning of the summer 1961, some boulders in the Ermine river bed were marked with paint (Fig. 94). Some of these boulders were moved during the flood periods; a few were found again and the distances they had travelled were recorded.

Some 45 samples of sand and gravel were collected in various river beds. There was usually a wide variety of deposits present at a site. It was therefore very difficult to get a sample which was really representative, e.g. for a specific section and/or a particular stage and course

Fig. 93: Ermine River bed 1961-1963, scour and fill at cross-section 3.

of the river. An attempt to collect at each sampling site material that appeared to be the coarsest or the finest did not solve this problem. Only random sampling on a large scale might give reasonably representative values.

b) Floods after Rainfalls

After a rainfall, the bulk of the water cannot percolate because of the negative glacier ice temperatures and the presence of continuous permafrost in the ground. As, in addition, the vegetation is sparse and the evaporation relatively small, practically all the rain water runs off almost immediately, collecting in the various runnels and stream channels causing brief but sharp peaked rises in the flow of these rivers (Figs 15, 92a) and consequently decisive geomorphological effects in the river beds (Rudberg, 1963b, p. 146, 148; Adams, 1963, p. 52-53; 1966, p. 63, 65, 67-68). Adams shows that after a concentrated rainfall the peak discharge of the glacier-marginal Ermine River considerably exceeded the maximum flow recorded at times of most intensive ice melt. The destruction of Adams' gauging installations in the Ermine River is just one example of the devastating effect of such floods (Figs 92a, b).

The successive profiles of cross-section 3 assembled in Figure 93 show that the greatest changes at this site, which is situated in a river reach of some 10^0 slope, were caused by three short, violent flood waters in July 1961, with approximate peak discharges of $19.5 \text{ m}^3/\text{sec.}$, $30 \text{ m}^3/\text{sec.}$, and over $50 \text{ m}^3/\text{sec.}$ respectively (Adams, 1966, p. 63, 65) and with estimated peak velocities of more than 5 m/sec. The modifications in the river bed during these few days were much greater than during the rest of 1961, or during the whole of the summers of 1962 and 1963, when no major rain floods occurred, and this despite the fact that in 1962 the total net ablation - which delivers the major part of the water running through the Ermine River channel (Adams, 1966, p. 71) - was several times larger than in 1961 (Müller, 1963e, p. 40).

Changes in the position of marked boulders corroborate the observations at the cross-sections. The selected photographs in Figure 94 show that it was exclusively during the short periods of flood discharge in July 1961 that most of these boulders were overturned or moved down-river, whereas during the second part of the summer 1961 and during 1962, 1963 and 1964 only minor modifications occurred such as the deposition of small sized material around boulder No. 5.

The changes caused by the three floods in July 1961 were of differing magnitude, the last one having the most decisive effect on the river bed (Fig. 93, period 2). Of course, the peak discharge and velocity were estimated to have been considerably larger on 26/27 July than on 5 or 13 July (see above), and this might account for the differences in geomorphological effect. On the other hand, it is difficult to assess the exact amount of water that passed through the profile site at each phase of the floods because at times part of the flood waters by-passed the meander loop where cross-section 3 was situated, using a new short route that was cut into consolidated moraine during the flood of 13 July (Fig. 93, situation sketch map). Nevertheless, according to laboratory and field data of various authors regarding the critical erosion velocity or shear stress (Hjulström, 1935, p. 298; Sundborg, 1956, p. 177; Fahnestock, 1963, p. A29, 31; Leopold et al., 1964, p. 170), the erosional impact exerted by the highly turbulent waters was theoretically large enough to initiate movement of most of the particle sizes present at the cross-section and during the first two floods if the bed material were truly unconsolidated. Especially at the time of the first flood, the bed surface was still frozen and remnants of river ice cemented the particles, thus increasing the resistance to erosion (5 below). Also, there was a relatively large amount of smaller-sized material packed around the boulders. Only by the time of the third flood was the coarse bed material sufficiently loosened and exposed to be successfully attacked by the flood waters.

The main scouring activity during the periods of highest discharge was in the middle and the left part of the bed (Fig. 93, period 1 and 2), which suggests that there was a shift of the main current towards the left, i.e. from the outer edge of the meander bend towards the convex bank. This would indicate - and photographs of the river in flood seem to confirm it - that the channel width, the mean radius of curvature and very likely also the wave length of the meander increased at times of highest discharge, whereas the amplitude probably decreased (Hjulström, 1949, p. 84-85; Wundt, 1953, p. 96; Leopold et al., 1964, p. 296; Schumm, 1965, p. 790). The modifications caused by the flood waters are usually not uniform over the whole section (Rudolph, 1962, p. 240); there can be net fill on one side and net scour on the other (Fig. 93, period 1). Again, the same flood can cause net scour in one reach of the river channel - usually a relatively steep one (e.g. at cross-



Figs 94a, b & c:
Marked boulders in Ermine
River bed:

a) during flood of 13/14 July 1961
(14 July 1961, 0600).



b) after flood of 13/14 July 1961
(19 July 1961, 1700), boulders
7, 10, 11, 12 have moved.



c) 22 August 1963, at end of
melt season, boulders 5
and 9 still at old place, new
small grained material
deposited between 5 and 9.
For scale, man in middle
ground.

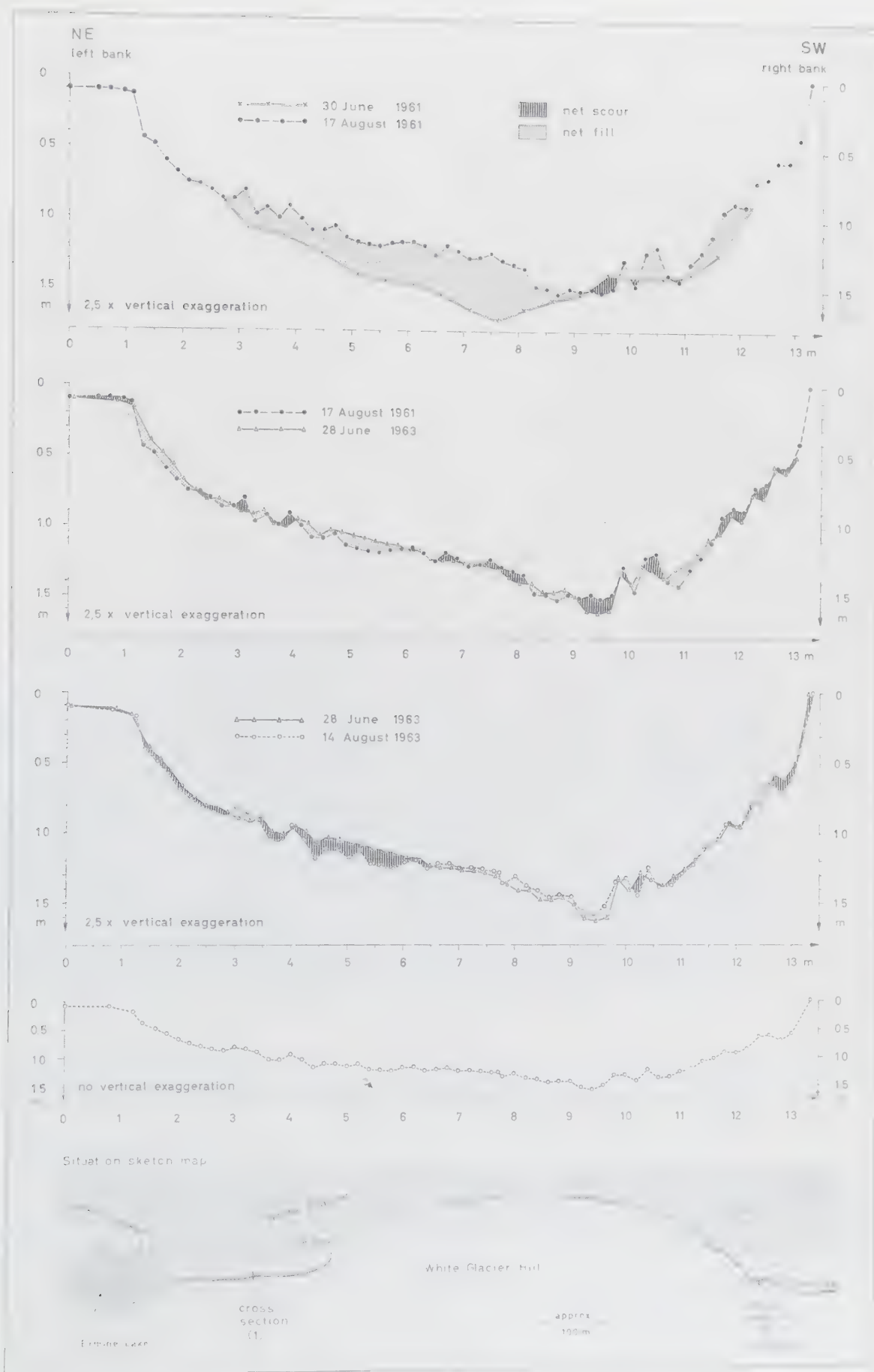


Fig. 95. Exmoor River bed 1961-1963. scour and fill at cross-section 1.

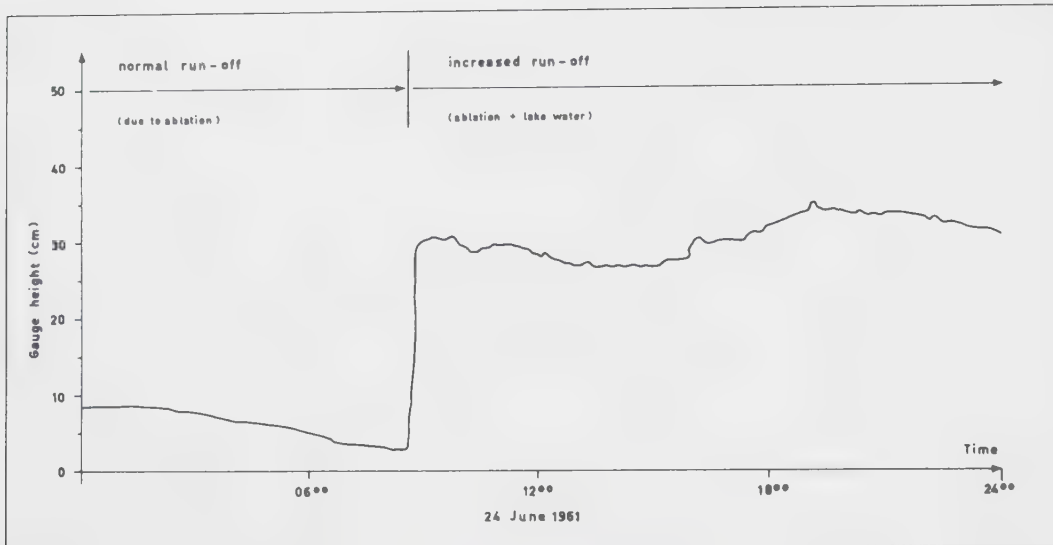


Fig. 96: Stage of Ermine River during drainage of small ice-dammed lake.

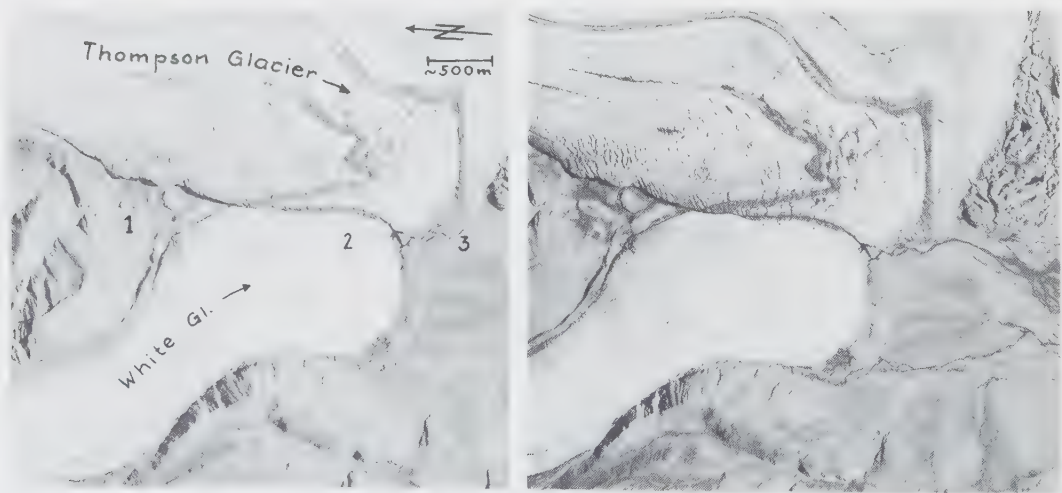


Fig. 97: Vertical air photos (stereopair, part of RCAF A 16 864 - 36/37, 13 August 1959): 1) empty Between Lake basin, 2) ice channel cut by outflowing Between Lake water, 3) Between River.

section 3, Fig. 93) - and net fill in another, usually a less inclined stretch (e.g. at cross-section 1, Fig. 95).

The coarse material in the Ermine river bed was moved during the floods in a discontinuous manner and over varying, but mostly short, distances (Tricart, 1962, p. 298; Fahnestock, 1963, p. A27; Leopold et al., 1964, p. 328). A marked gabbro boulder, approximately 120 x 70 x 70 cm³ in size and over 1000 kg in weight, was moved some 200 m downstream during the second and third floods of 1961, and was found to be still in that position in August 1963. Another boulder, with an estimated weight of some 300 kg, was transported some 450 m, whereas one just about half that size travelled only some 80 m. Others did not move at all (boulders Nos 5 and 9, Fig. 94), or were just rocked and overturned (boulder No. 3, Fig. 94). The course of boulder No. 8 (700 kg) seems to be typical for the transport pattern of the Ermine River. During the first flood of 5 July it did not move at all, but some smaller particles around it were removed and some ice was melted. On 13 July the flood waters managed to roll this boulder about 4 metres (Fig. 94b). The flood of 26/27 July 1961 moved it a further 100 m downstream and partly covered it with coarse material. During the remainder of 1961 and throughout 1962, 1963, and probably 1964, it remained there. In the course of some 4 years, therefore, this boulder was in motion only for a few hours. Another example of the discontinuity of movement of the coarse bed material and the brief transport time is shown in Figure 93. During the second flood of 1961 some coarse material was moved into the profile and deposited in the right hand side of the river bed. It remained there for some two weeks; the third flood of 1961 partly removed it. The rest - as well as some additional bed material - was carried off during the peak of the 1962 ablation season. In the summer of 1963, on the other hand, some fine grained material was deposited in this part of the Ermine River.

At cross-sections 2 and 3 and in the relatively steep reach above and below, it was observed that after a flood there was a deficiency in small grain sizes suggesting they were flushed out during the flood (Moosbrugger, 1954, p. 244; Wundt, 1962, p. 203; Tricart, 1961, p. 9), whereas the material that was carried into this stretch was mainly coarse grained. Later on, at lower flow, finer material was again deposited (Fig. 94c), levelling out to some extent the bed (Leopold et al., 1964, p. 84).

The present study shows that only during times of flood discharge is the Ermine River competent to move the coarser bed material. As a surprising number of large boulders is found in the Ermine River bed and in other rivers in the Expedition Area and as many of these boulders show varying degrees of edge-rounding and are often not lichen-covered, it is suggested that already in earlier summers high waters comparable to the ones experienced in the summer of 1961, and with similar effects, must have occurred. Rudberg (1963a, p. 219) thinks that rain storms causing such floods do not occur very frequently as he observed networks of ice-wedge polygons in the lower part of alluvial fans. On the other hand, Müller (1963d, p. 35) after investigating snow, firn and ice layers in a deep shaft on the McGill Ice Cap, suggests that "the 1961 summer, which was thought to be exceptionally wet, was in many respects quite an average one", and that - during the last 35 years - similar "wet" summers may have occurred several times. Jackson (1961, p. 141) reports that "occasional exceptionally heavy (rain) showers" in the Queen Elizabeth Islands are not uncommon.

At all observed sites in the Ermine River channel, the bed-shaping periods in 1961 were just the few hours of peak discharge due to concentrated rain showers, especially that of 26/27 July. The discharge of this river during these few hours amounted to 15% of the total for that summer (Adams, 1966, p. 71, Table 19).

In regions with temperate glaciers where the retention qualities of the glacier and of its surroundings are greater, there is normally a less direct effect of a rainfall on the river discharge and therefore a less concentrated geomorphological effect.

c) Floods Caused by Outflowing Lake Waters

In the Expedition Area it has been observed that the flood waters from draining ice-dammed lakes can cause decisive changes in the beds of marginal and proglacial rivers. There are many historical and modern reports of the drainage of ice-dammed lakes and the devastating effects of their flood waters; most of these, however, are from regions with temperate glaciers.

The trace (Fig. 96) recorded by an automatic level recorder, installed by W.P. Adams

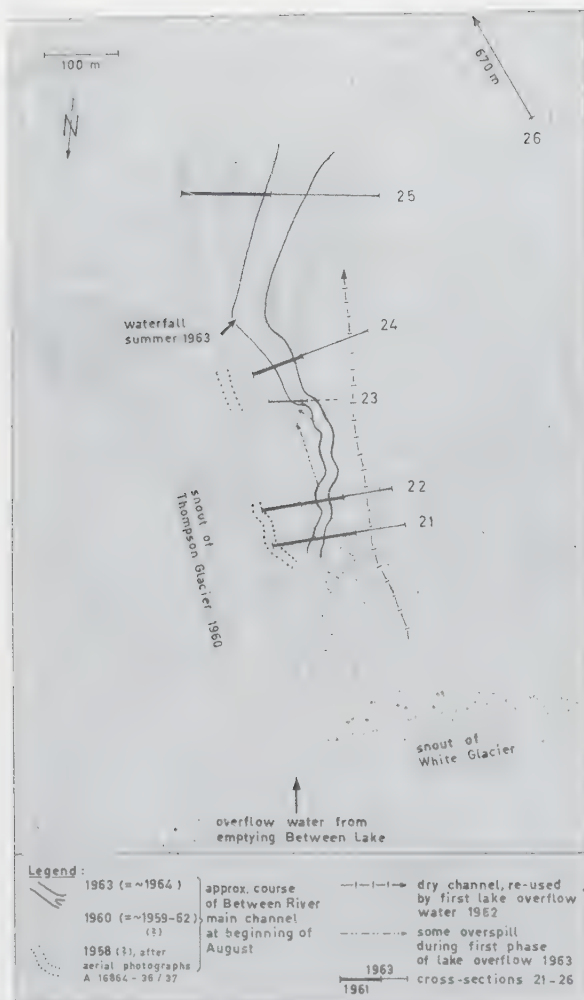


Fig. 98a: Between River main channel 1958(?) - 1964.

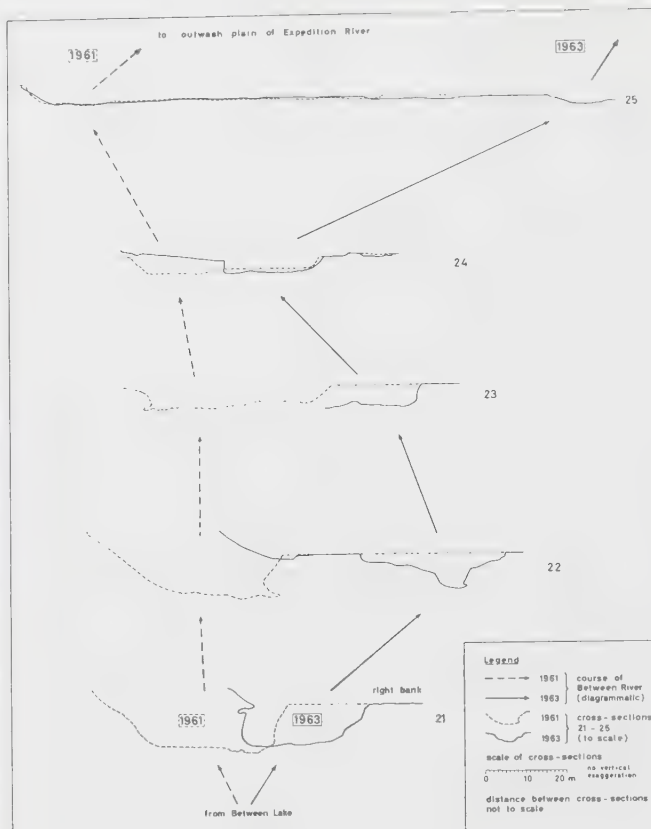


Fig. 98b: Between River, 1961-1963.



Fig. 99: Between River (10 August 1963, 1600) looking upstream. Foreground: rest of old river bed and new river bed of summer 1963. Background: White Glacier with old and new overflow channels.



Fig. 100: Change at cross-section 21, Between River bed, summer 1961-1963.

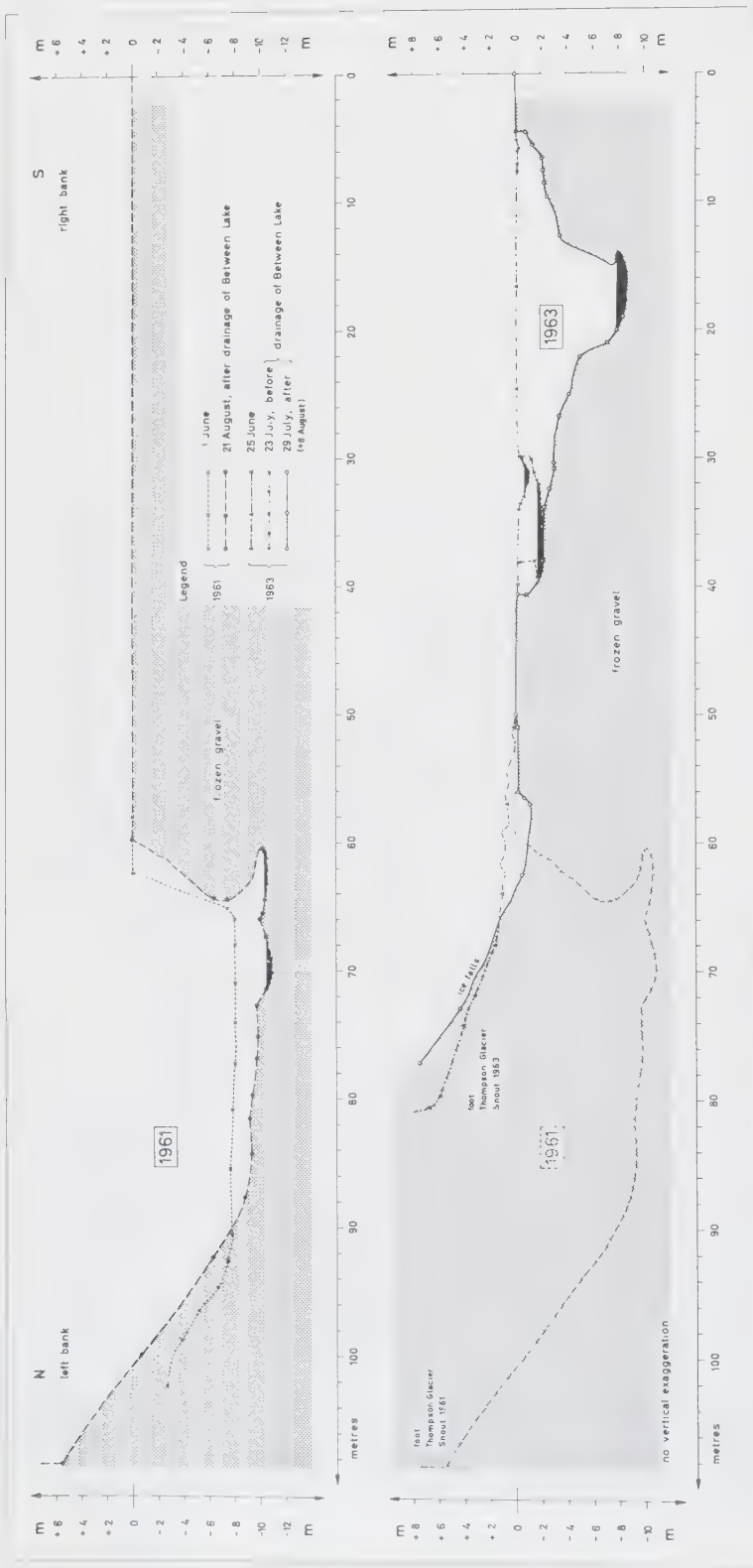


Fig. 101: Change at cross-section 22, Between River bed, summer 1961-1963.

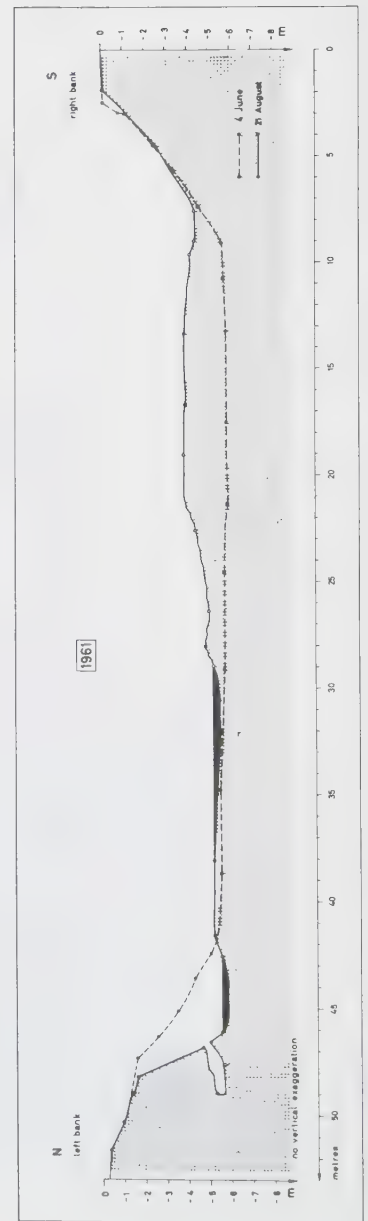


Fig. 102: Change at cross-section 23, Between River bed, summer 1961.

Fig. 103: Change at cross-section 24, Between River bed, summer 1961-1963.

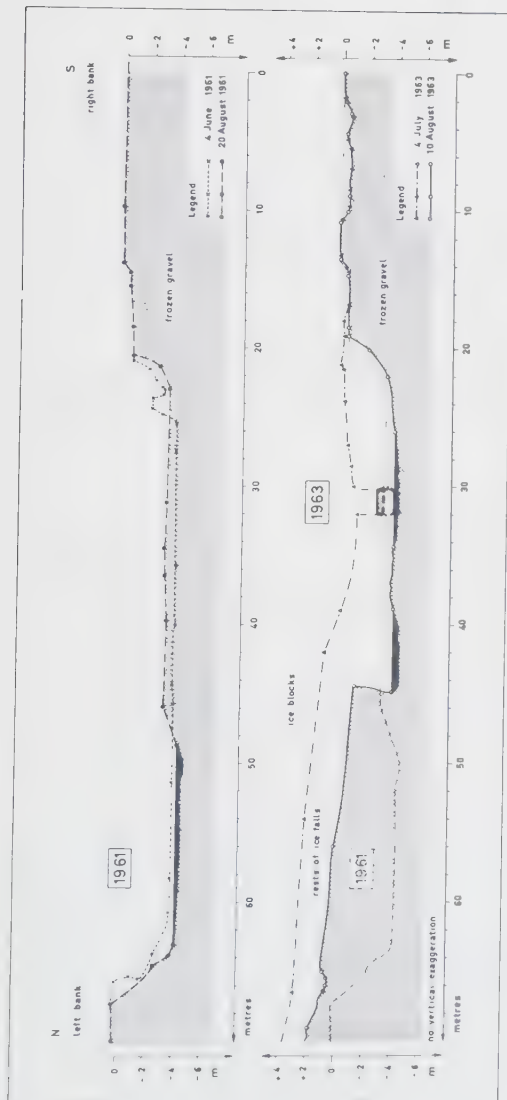
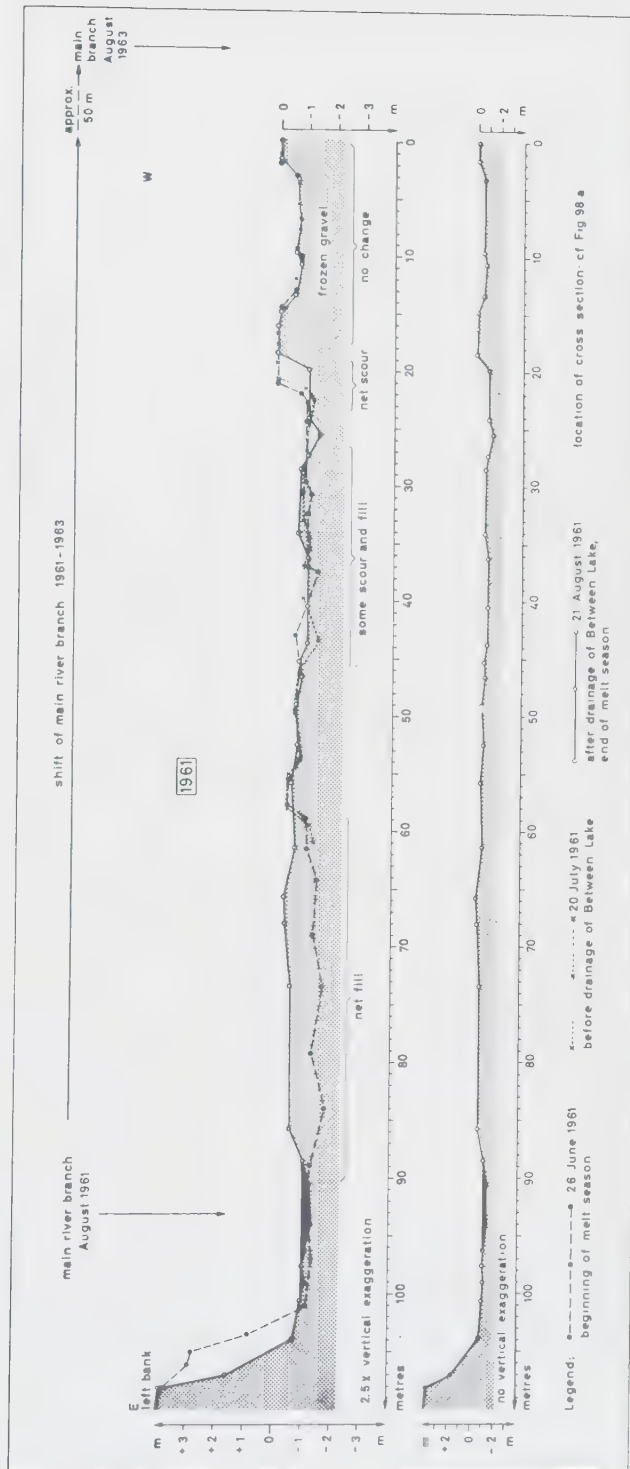


Fig. 104: Change at cross-section 25, Between River bed, summer 1961 (+ shift of main branch 1961-1963).



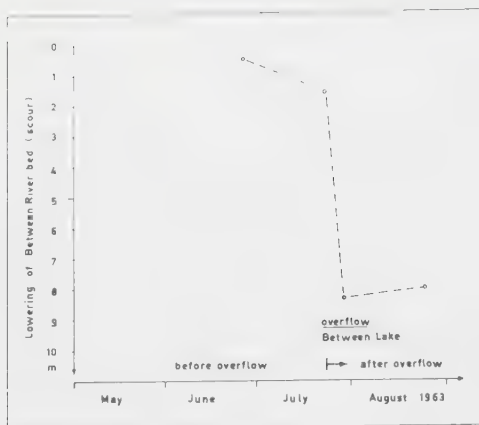


Fig. 105: Mean elevation of Between River bed floor at cross-section 22 during summer 1963.

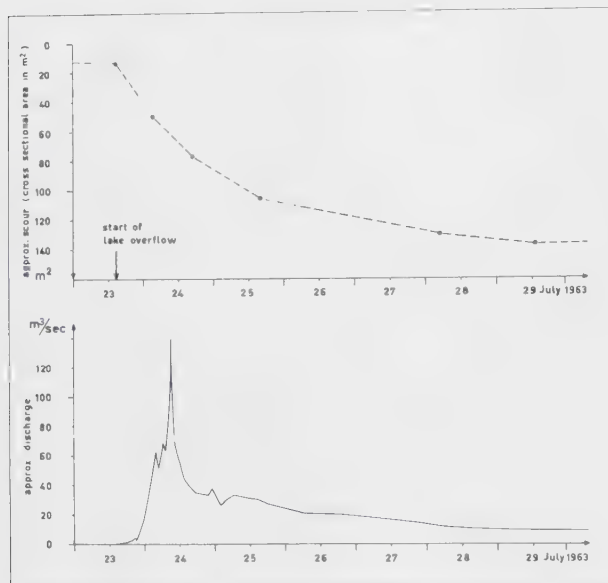
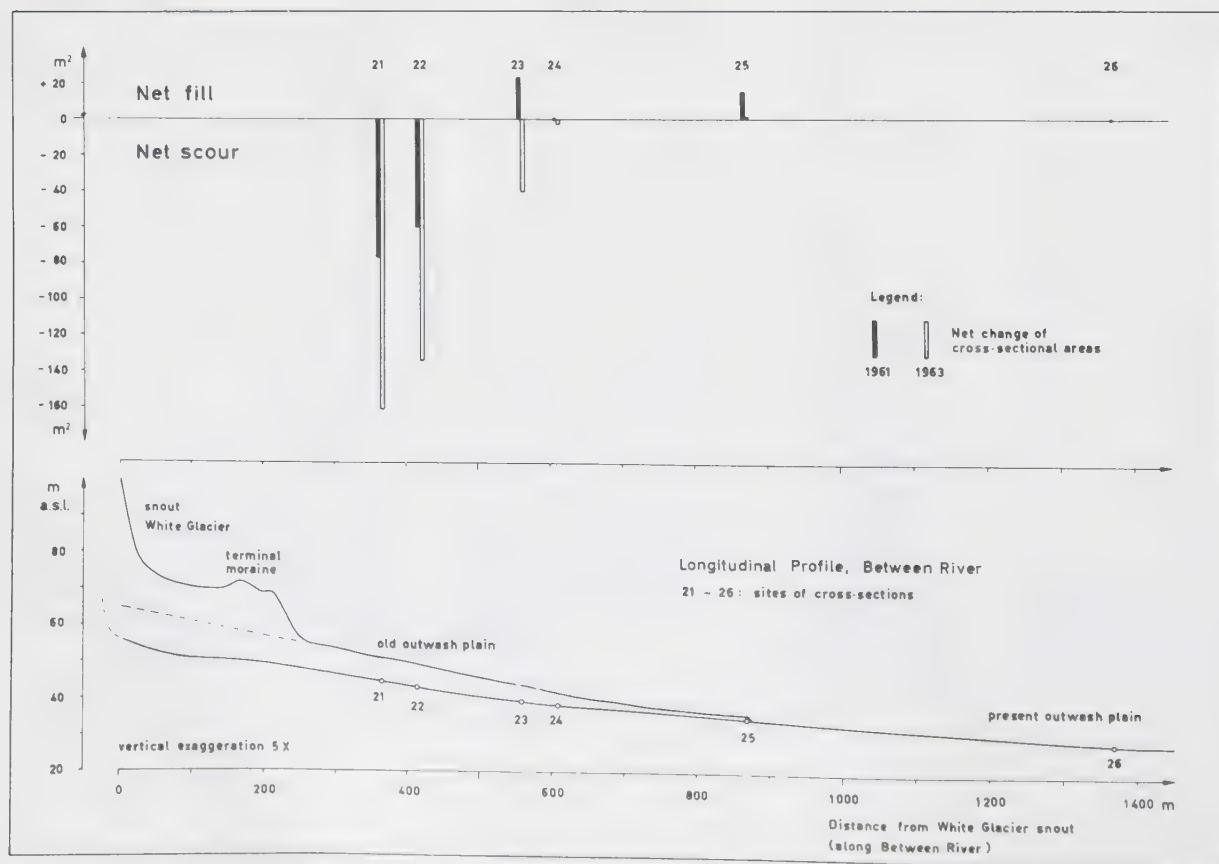


Fig. 106: Scour at cross-section 22, Between River, during outflow of Between Lake, summer 1963.

Fig. 107: Net change of cross-sectional areas, Between River 1961, 1963.



in the Ermine River, shows the typical development of the discharge during the outflow of a small one of the many ice-dammed lakes in the Expedition Area. A fast rise to peak discharge is followed by a period of high stage which decreases slowly. So, in contrast to the short, but very peaked, flood waves rushing through the river beds after rain storms, in the case of the drainage of an ice-dammed lake high discharge is maintained usually over a longer period.

The effect of such floods was studied through the modifications in the Between River bed caused by the outflow of Between Lake. The longitudinal profile (Fig. 37) shows the course of the outflowing lake water which is running in a deep ice channel (cf. Chapter B) across the tongue of the White Glacier. Leaving the glacier, it forms the Between River which is cutting through the terminal moraine and into the frozen gravel of an old outwash fan but on reaching the less inclined modern outwash plain of the Expedition River it becomes braided. The estimated peak discharges of the Between River during the outflow of Between Lake were $140 \text{ m}^3/\text{sec.}$ in 1963 and $40 \text{ m}^3/\text{sec.}$ in 1961 (not counting the two very short maxima of some $50 \text{ m}^3/\text{sec.}$ occurring towards the end of the drainage period of 1961 after the outlet channel had been temporarily blocked).

In Figures 98a, b and 99 some progressive changes in the course of the Between River are shown. The Between River, which is a proglacial river of the White Glacier but which is marginal with respect to the Thompson Glacier snout, could use the same bed only for a very few years because of the advancing Thompson Glacier. The glacier cliff advances some 20 m per year (Müller, 1963g, p. 75), continuously narrowing and finally covering the river bed so that the waters have to dig a new channel every 3 - 4 years; this last occurred in 1963.

Changes in the shape of the Between River bed during the summers of 1961 and 1963 are shown in Figures 98b and 100 to 104. It was during the floods caused by the drainage of Between Lake that the largest changes were observed, whereas in the periods before and after only relatively minor modifications occurred. At cross-section 22 the remarkable deepening and widening of the river bed in 1963 (Fig. 101) occurred almost exclusively during the few days of flood discharge (Fig. 105). The effects in 1961 were less spectacular, but again the most decisive channel shaping was done during the outflow of Between Lake as can be seen at cross-section 21 (Fig. 100). In the summer of 1964, when Between Lake did not drain, only very small modifications took place (personal communication and photos, C.S.L. Ommanney).

Cross-sections 21 and 22 both lie in the reach where scour is usually greater than fill. But also in river stretches where fill exceeds scour, e.g. at cross-section 25 (Fig. 104), the greatest modifications to the channel can be correlated to the few flood days. The amount of undercutting and the speed of recession of the river banks also varies markedly with the discharge.

That the channel changes in 1963 were much larger than those of 1961 (Fig. 107) was, in the first place, due to the fact that the Between Lake drained much faster and with larger peak discharges (cf. Chapter B). In addition, in 1963 the Between River had to dig a new channel between the cross-sections 21 and 24 (Figs 98, 99) because the old one was blocked by ice falls from the Thompson Glacier cliff. The slope of the new channel was at first quite steep as the water was dammed up about 10 m above the level of the old bed and detoured the ice fall on top of an old outwash plain. From this elevated surface, the Between River cut a channel into the frozen gravel until it was again on a level with the old river bed which it joined further downstream. In 1961, on the other hand, the river could use a channel with a slope approximately at grade and a cross-section already shaped by previous floods to accommodate high discharges.

The largest modifications occur during the first part of the flood, that is during the time of rapidly increasing and maximum discharge (Fig. 106). Figure 108 shows the 1963 Between River near cross-section 22 shortly before peak discharge and one day later. It appears that the upper part of the final shape of this cross-section was cut during the few hours of increase to full and overflow discharge and that the lower part of the channel was cut after the peak of the flood by the enchannelled decreasing waters. Towards the end of the flood the further deepening of the channel at cross-section 22 proceeded much more slowly and the left part of the bed was even filled up a little. Such deposition towards the end of the flood was also observed in 1961, e.g. on the left side of cross-sections 23 and 24 (Figs 102, 103).

Comparing the amounts of net scour and net fill in the different cross-sections (Fig. 107)

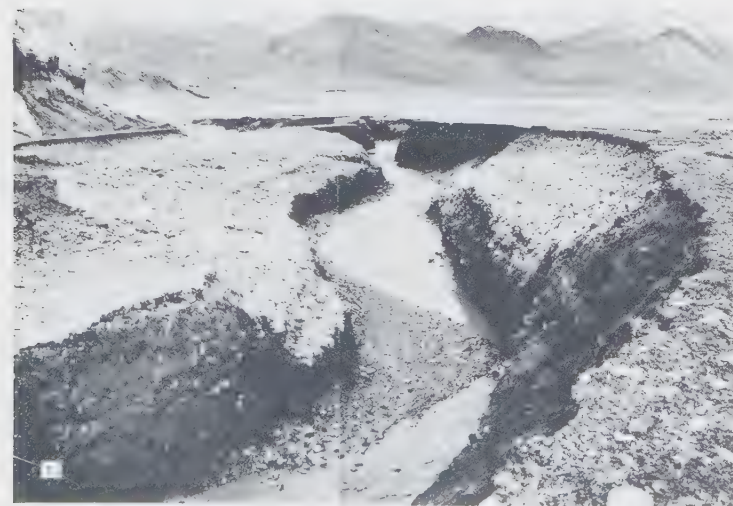


Figs 108a, b & c:
Between River at cross-section 22,
looking downriver:

a) 24 July 1963, 2300, first day of
flood discharge (approx. $60 \text{ m}^3/\text{sec}$)
because of drainage of
Between Lake, shortly before
peak discharge.



b) 25 July 1963, 1600, end of
second day of flood discharge
(approx. $30 \text{ m}^3/\text{sec}$), stream
in foreground some 11 m wide.



c) 12 August 1963, 1700, low
water after flood discharge,
river meanders.

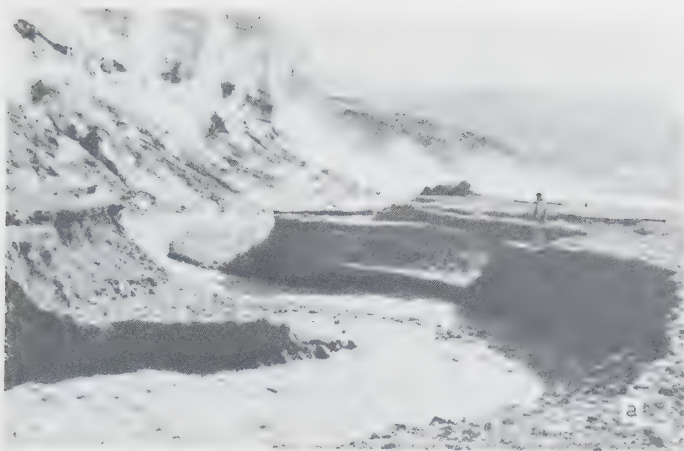


Fig. 109a:
Between River near cross-section 24
shortly before Between Lake overflow
(23 July 1963, 1200).

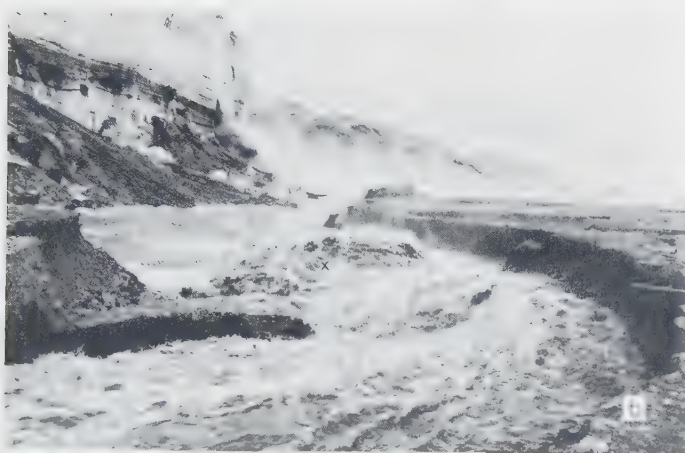


Fig. 109b:
The "Umlaufberg" (x) is being
carried off during flood (24 July
1963, 0400).

Table 10: Some data on suspension and solution load, Between River, 1961 and 1963

Sampling			Dissolved substances		Suspended material		Approx. discharge m ³ /sec	Remarks
Date	Time	Place	Total (mg/l)	Organic (% of total)	Total (mg/l)	> 63 μ (% of total)		
1961 26 June	--	some 400 m from snout	52	26.9	1333	~2	~ 0.5	still much river ice in river bed
12 July	1900	"	292	21.9	1769	~1	(~5*)	*river dammed by ice-fall
22 July	2345	"	112	18.8	3662	6	~ 8	2 $\frac{1}{2}$ hours before start of overflow of Between Lake; much rain dur- ing the preceding days
23 July	1900	"	182	24.7	2575	8	20	17 hours after start of overflow of Between Lake (6 hours before peak discharge)
24 July	1700	"	207	28.0	1929	18	30	39 hours after start of overflow of Between Lake (16 hours after peak discharge)
21 Aug.	1830	"	281	14.2	701	1.5	1	end of melting season; several days after the emptying of Between Lake was completed
1963 24 July	1730	"	63	3.2	730	46	35	24 hours after start of overflow of Between Lake (3 hours after peak discharge of approximately 140 m ³ /sec.)
13 Aug.	0430	at snout	178	2.2	812	2.5	1	several days after emptying of Between Lake was completed

one sees that scour decreases very rapidly with increasing distance from the glacier snout and decreasing slope of the bed. At a distance of some 600 m from the snout, in summer 1961, scour already equalled fill. This means that the morainic material and the gravel eroded within this reach is carried during floods over only a relatively short distance and deposited again as already observed for the Ermine River.

The amount of scour and fill in the different parts of a cross-section can vary greatly. The river can scour in the right part and fill in the left at the same time. Likewise a site which undergoes scour may, a few days later, receive fill. Unfortunately, no surveying was possible during flood periods. But a series of photographs showing the shifting of the main current and the development of new bars gives valuable indications of what happened during time of high discharge. In cross-section 21 (Fig. 100) there was some scour in the left part of the river bed and some deposition on the right side just before the outflow of Between Lake and in the first phase of the flood. Later on, the main scour occurred on the right side of the river bed, whereas on the left side a gravel bar developed. A similar development was observed at cross-section 23 (Fig. 102) where vertical scour and undercutting were taking place on the right side of the bed during the first part of the flood, while later on the main current eroded the left part and considerable deposition occurred in the middle and on the right. At cross-section 24 (Fig. 103) in the earlier part of the 1961 flood a slumped piece of the river bank was carried off by the strong waters that rushed through the right part of the river bed. At the end of the flood period the river occupied the left side of the cross-section, scouring a little the foot of the left bank, whereas in the right part of the river bed considerable amounts of gravel and sand were deposited.

This alternation of deposition and scour at one site reflects shifts of the main current. A study of both the photographs and the surveyed cross-sections suggests considerable changes in the flow pattern throughout the summers and from year to year. In general, during low stage of the river a shortening of the meander length was observed. In Figure 108c the river course shows the long meanders of the stream at flood discharge, at low stage the meanders have a much smaller wave length. During flood discharge the wave length of the Between River increased (Fig. 109) as observed for the Ermine River. Downstream migration of bends and the cut-off of whole meander loops also took place.

Where the Between River leaves the glacier snout almost no suspended load was found in the water whereas at site 22, after a run of some 400 m through the terminal moraine and outwash gravel, the river had already picked up considerable amounts of suspended load (Table 10) so that the concentration was approximately of the same order as reported from glacial rivers elsewhere. The concentrations varied much from sample to sample. How far the primitive sampling techniques are responsible for this or how far the differences were due to the normal fluctuations of the suspended load within the cross-section, with time, discharge and turbulence is not known. Nevertheless, it seems that, at least in 1961, the water contained higher amounts of suspended particles, especially of coarse ones, during periods of high discharge than at low stage (Table 10), although the increase was not as marked as reported from lake floods elsewhere (Collet, 1925, p. 30; Thorarinsson, 1939a, p. 210-211). The data are too few to ascertain if the peak of sediment concentration in the Between River preceded the peak of discharge as has been reported elsewhere (Lütschg, 1926, p. 269; Hjulström, 1935, p. 411; Sundborg, 1956, p. 300-301; Leopold et al., 1964, p. 71; Douglas, 1964, p. 463). It is interesting to note that in 1961, when the discharge was lower than in 1963, the measured concentrations of load were larger (Table 10).

A comparison of suspended material collected from the Between River at the site of cross-section 22, before, during and after the lake emptyings of 1961 and 1963, shows that during high discharge considerably coarser particles were in suspension than at low stage, which is to be expected because of the higher velocities and turbulence (Fig. 110). The suspended particles collected during the height of the flood of 1963 were coarser than the ones in the 1961 flood where the discharge was smaller. The largest particles in these surface samples had a diameter of some 1.5 mm. Even larger particles must have been in suspension at greater depth.

To assess the solution load present in the Between River some water samples were analysed (Table 10). The concentrations are larger than those reported from glacial rivers in Alaska, in the Antarctic, or in temperate regions (Huber, 1950, p. 45-80, summer run-off; Rainwater and Guy, 1961, p. C-2-4; Keller and Reesman, 1963, p. 65; Stenborg, 1965, p. 150), but they are approximately of the same order of magnitude as those found by Thorarinsson (1939a, p. 210-211, 213).

Fig. 110: Suspended load in Between River, near cross-section 22, summer 1963: before (27 June), during (25 July), and after (9 August) drainage of Between Lake.

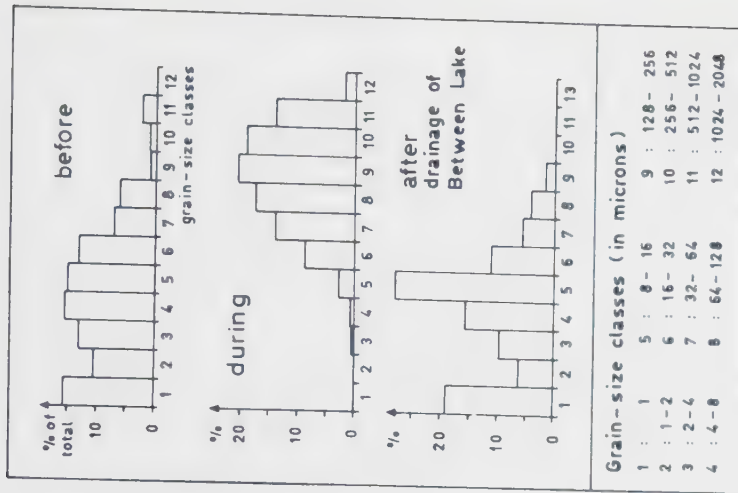
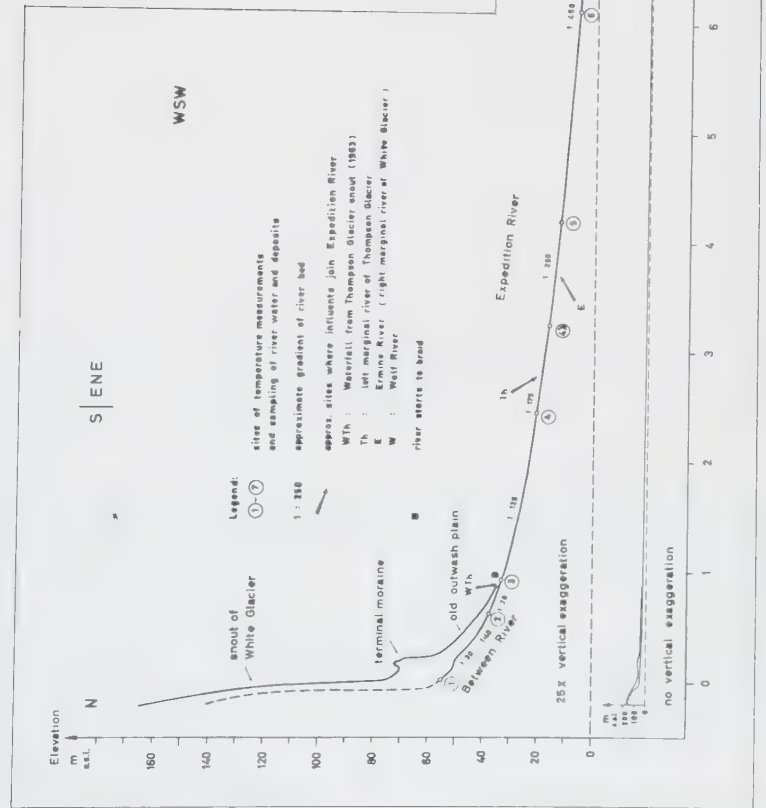


Fig. 111: Longitudinal profile Between River bed - Expedition River outwash plain.



in Icelandic glacial rivers, or by Huber (1950, p. 45-90) in Swiss ones during the winter months. A direct comparison of the various data is only of limited value as the environmental factors influencing the amount and composition of the dissolved load differ.

The solution load in the Between River obviously varies greatly during one summer and from year to year. Due to the scarcity of data final conclusions on these variations are not possible and, as yet, the relationship between dissolved substances and discharge is not fully understood (Douglas, 1964, p. 462). However, some comments can be made. During times of high discharge resulting from the outflow of Between Lake the concentration of dissolved solids in the Between River was lower than before and after, except for the period very early in the melt season when there was still much river ice in the bed preventing the water from making contact with the bed material. The data suggest, too, that in 1963, at the peak of the flood caused by the outflow of Between Lake, the concentration of dissolved substances was lower than in 1961 as was observed also for suspended matter. This might be due to the fact that the sampling in 1963 took place only some 3 hours after the peak river discharge which was much larger than that of 1961. The Between Lake water of 1963 seems to have been more dilute than that of 1961 (Table 8). More organic substances were present in the flood water in 1961 (Table 10). The relatively high concentration of 12 July 1961 is probably a special case as prior to that date the Between River had been dammed by an ice fall, and the water had more time to acquire soluble constituents from the river bed.

The chemical composition of the Between River water of 24 July 1963, that is at the height of the flood discharge, is very similar to that of the outflowing Between Lake water (Table 9), in spite of the fact that it has passed through an ice channel of some 1300 m length to the tongue of the glacier and then some 400 m through a bed cut into moraines and gravels. It is interesting, however, that on 13 August 1963, at low stage and after the Between Lake had drained, the water collected directly at the ice front, where the Between River leaves the White Glacier snout, contained a much larger amount of dissolved solids, especially of sulphates (Table 9). As previously suggested this might indicate a subglacial contact with gypsum.

The few data on changes in the river bed cross-sections and on concentrations of suspended and dissolved material allow only a very rough estimate of the total erosional effect of the water of the draining Between Lake and of the total load moved through the Between River bed (Table 11). The relative amounts of bed load, suspension load, and solution load varied during the flood. Towards the end of the drainage of Between Lake relatively more suspended and dissolved material was transported through the investigated reach than at the peak of the flood and only minor changes in the shape of the river bed took place. The bed-shaping work was done almost exclusively during the few days of peak flood (Fig. 106).

The outflowing Between Lake water contained no, or very little, bed load and only minor amounts of suspended load. Therefore most of the coarser material moving through the Between River bed had been acquired after leaving the White Glacier tongue in the few hundred metres where the river is eroding the terminal moraine and the old outwash gravel. On the contrary, most of the dissolved substances present in the Between River water samples collected in 1961 and 1963 during the period of the Between Lake drainage, originated from the lake and its tributaries with only minor additions from the short below-glacier stretch. Most of the coarse material was deposited again after a relatively short distance of transport, whereas the dissolved substances and much of the finer suspended material moved on out of the investigated reach.

The Between River starts to branch near cross-section 25 with a slope of about 1:60. It takes on a braided pattern and joins the Thompson and Ermine rivers to form the Expedition River which flows to the fiord over an outwash plain of about 12 km length and 700 to 2000 m width. The longitudinal profile of the outwash plain (Fig. 111) resembles in shape an Icelandic sandur studied by Hjulström (1952, Fig. 4, p. 340). Usually there were only one or two main branches with numerous secondary ones, the pattern changing throughout the run-off season and from year to year (Figs 112-114). Bars developed similar to those described by Hjulström (1952, p. 340-341) and Krigström (1962, p. 331-339) for Icelandic sandurs. Some channels, cut off from the water supply, dried out, whereas in others the water volume increased, new channels were dug and old re-occupied ones deepened and enlarged.

The greatest changes occurred during high discharge. Not only the many small branches but also the main branches shifted considerably at times (Figs 113, 114). The photographs (Fig. 112)

Table 11: Estimates of the total amount of material eroded and/or moved by the Between River during the outflow of Between Lake in July 1961 and 1963.

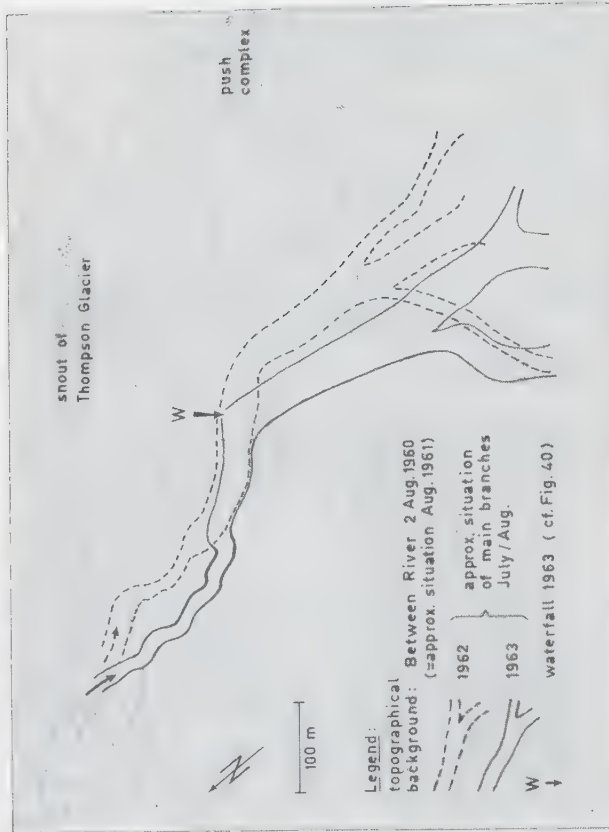
	1961	1963
Number of days until 2/3 of Between Lake had drained	4 days	2.25 days
Volume of water that flooded through Between River during this period	7 000 000 m ³	6 500 000 m ³
Peak discharge	40 m ³ /sec.	140 m ³ /sec.
Bed material (frozen gravel and moraine) eroded from the Between River bed (in a stretch of 600 m)	60 000 t	110 000 t
Material moved in suspension	17 500 t	6 500 t
Dissolved substances load	1 400 t	450 t
Average per m ³ floodwater: bed material	8.5 kg	17 kg
suspension load	2.5 kg	1 kg
solution load	0.2 kg	0.07 kg

taken from the snout of the Thompson Glacier show the outwash plain before and during the outflow of Between Lake. Even during the highest discharge not the whole of the outwash plain was covered by water, as was also observed by Krigström (1962, p. 329, 331, 341) for a sandur plain in Iceland. The water does not always follow the lowest part of the plain but prefers sometimes higher sections and may flow simultaneously in various branches on different levels (Hjulström, 1952, p. 338; Hoppe et al., 1959, p. 10, Fig. 10; Krigström, 1962, p. 341; Fahnestock, 1963, p. A50; Leopold et al., 1964, p. 289). This is illustrated by the cross-section 26 profile (Fig. 114) which has local differences in relief of up to 2 m and an overall elevation difference of some 7 m similar to cross-sections cited by Hjulström (1952, Fig. 5, p. 340) and Hoppe et al. (1959, Fig. 10, p. 10). The left part of cross-section 26 although lower than the right, remained almost unchanged throughout the summer of 1963, whereas the right section was modified considerably. Large parts of the outwash plain will, however, remain untouched by the water during the summer or perhaps even during many years (Leopold et al., 1964, p. 328). Although the longitudinal profile of the outwash plain (Fig. 111) appears as a smooth curve, in detail, the individual channels show small steps and variations in slope, as also observed by Fahnestock (1963, p. A51), but for a much steeper valley sandur on Mt. Rainier. That deep channels are often eroded in the proximal, steeper zone of a sandur, especially at the onset of the recession of a glacier, has been stated by various authors such as Troll (1926; 1957, p. 215) who calls them "Trompetentälchen"; Arnborg (1955b, p. 214); Okko (1956, p. 110) and Krigström (1962, p. 329). Similar deep channels have been cut by the Between River several metres into the now fossil uppermost part of the Between River-Expedition River outwash plain.

In 1961 a study was made of the changes in a small branch of the Expedition River at the right margin of the outwash plain (Fig. 115) which could be visited without danger throughout the whole summer. The data shows that on one single day, 6 August, an average fill of some 20 cm occurred whereas during the rest of the summer there were only minor or no modifications (Fig. 116). In 1963, however, the same location suffered considerable scour.

The lateral erosion observed in the Expedition River valley compares with the marginal erosion of a valley sandur reported by Krigström (1962, p. 339) from Iceland. In August 1961, for example, a main branch of the Expedition River undercut the margin of an old outwash plain, probably due to the outflow of an ice-dammed lake up-glacier (Beschel, 1961a, p. 189) and exposed a peat deposit.

The observation period was too short to ascertain whether the river is aggrading or degrading in the braided section. At cross-section 25 (Figs 104, 107) for instance, the observed net change differed from year to year: in 1961 and 1962 there was net fill within the section but in 1962 net scour in the extended western part, and in 1963 there was just about as much scour as fill. At cross-section 26 (Figs 107, 114) during the summer of 1963 fill in several parts of the profile about equalled the amount of scour that took place in others.



Figs 113a & b:

- a) Outwash plain of Between River - Expedition River. Thompson Glacier shout in background, 30 July 1961, 2100,
- b) Main branches of Between River, 1960, 1962, 1963.



Figs 112a & b:

- Outwash plain of Expedition River looking down valley (viewed from shout of Thompson Glacier). Dotted line indicates approximate position of cross-section 26 (some 700 m long).
- a) 6 July 1963, 1700, before flood,
- b) 25 July 1963, 1300, flood discharge resulting from overflow of Between Lake and strong ablation.

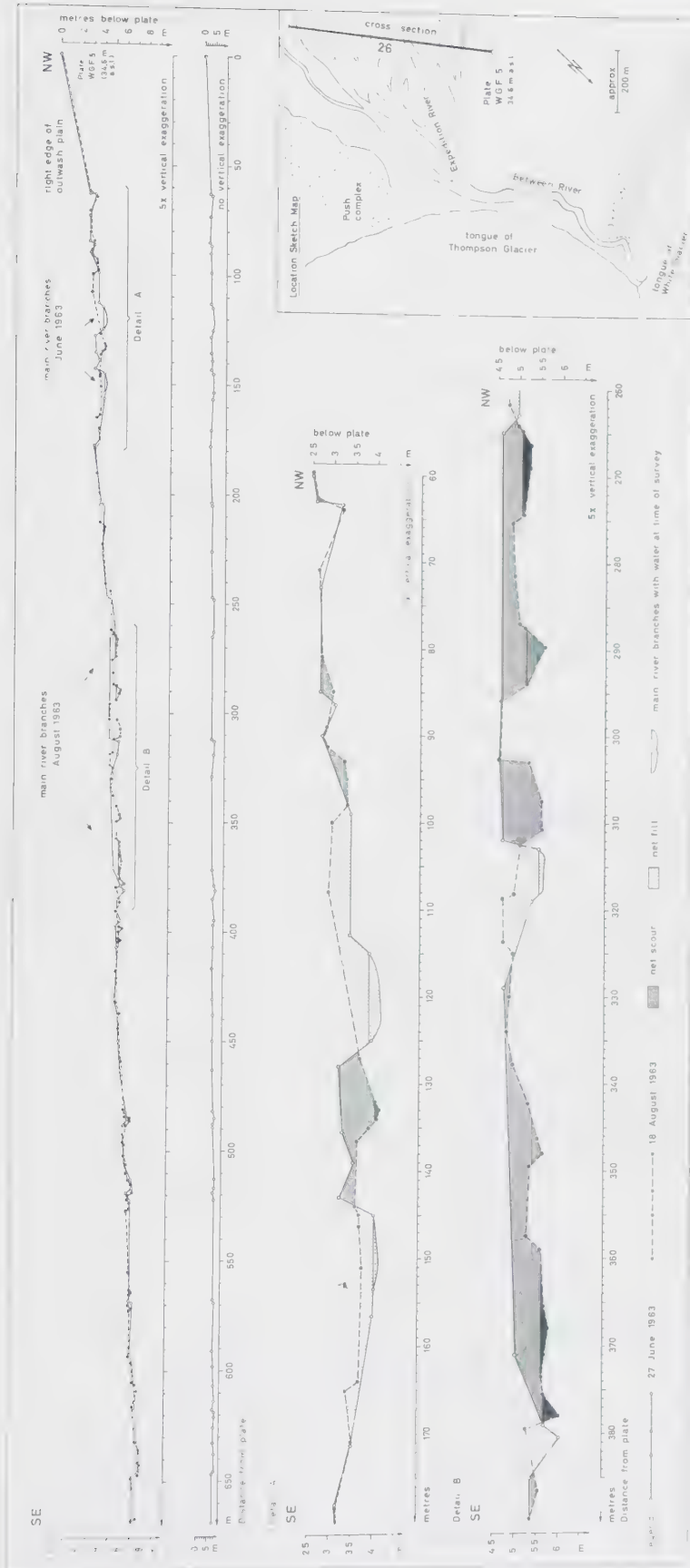


Fig. 114: Cross-section 26, Expedition River outwash plain, June and August 1963.

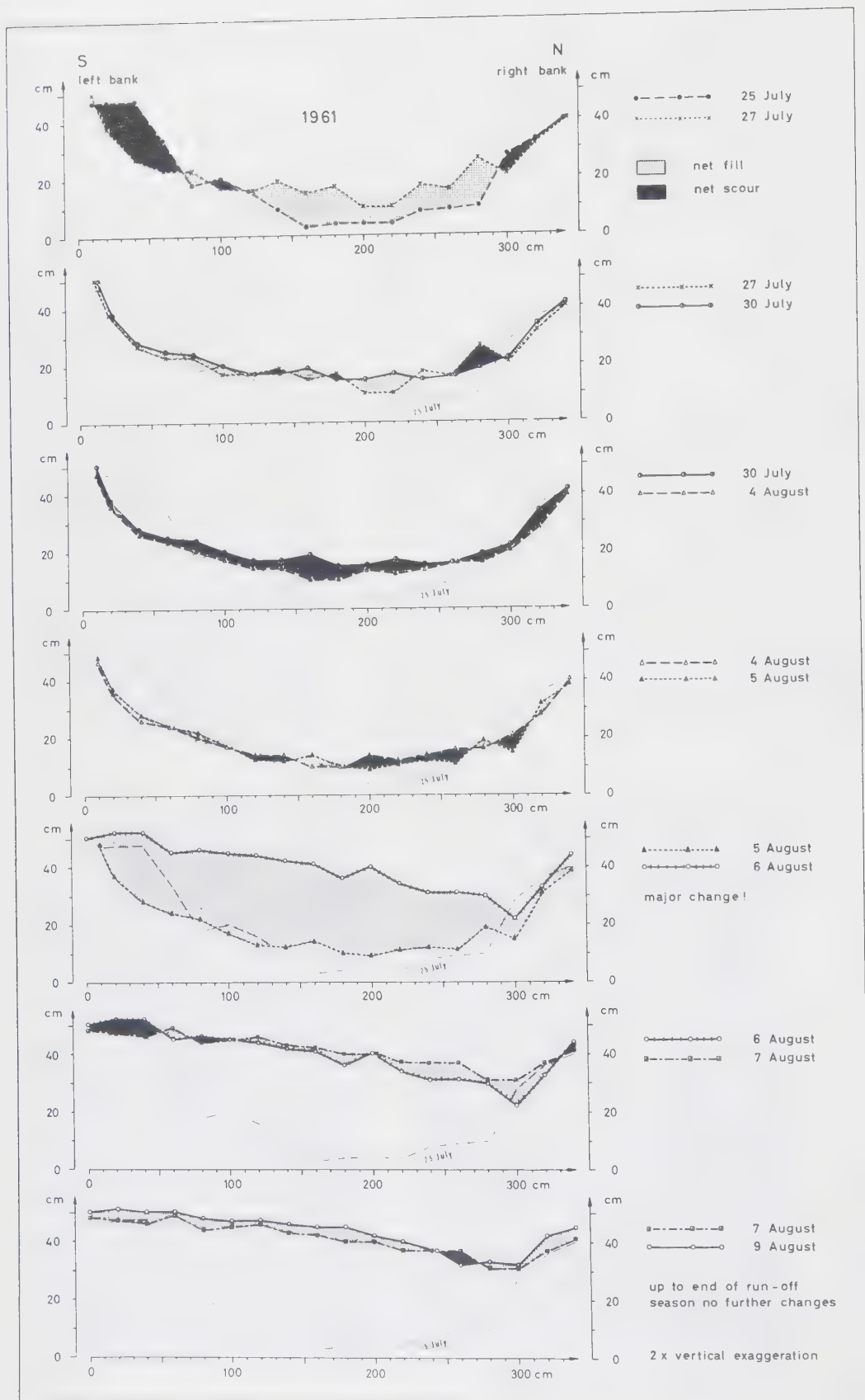


Fig. 115:
Channel cross-
section during
summer 1961,
small branch
of Expedition
River at right
edge of out-
wash plain.

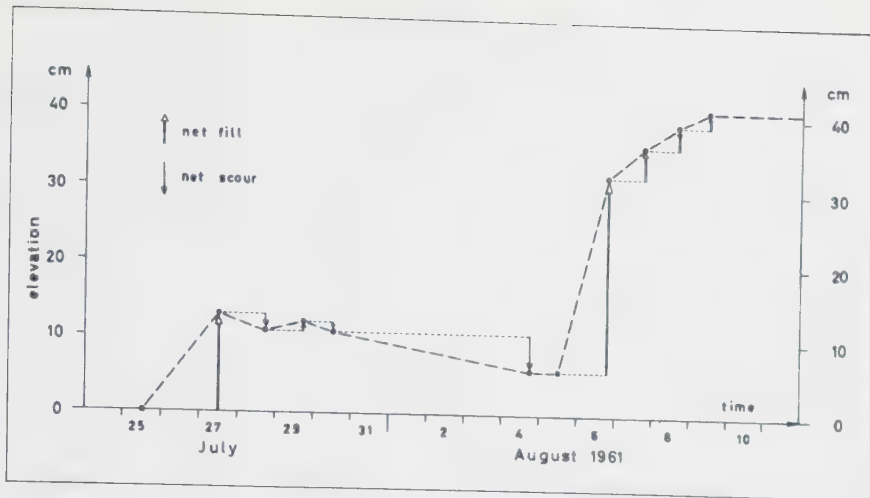


Fig. 116:
Elevation of point A in a small branch of Expedition River at right edge of outwash plain, July - August 1961 (cf. Fig. 115).

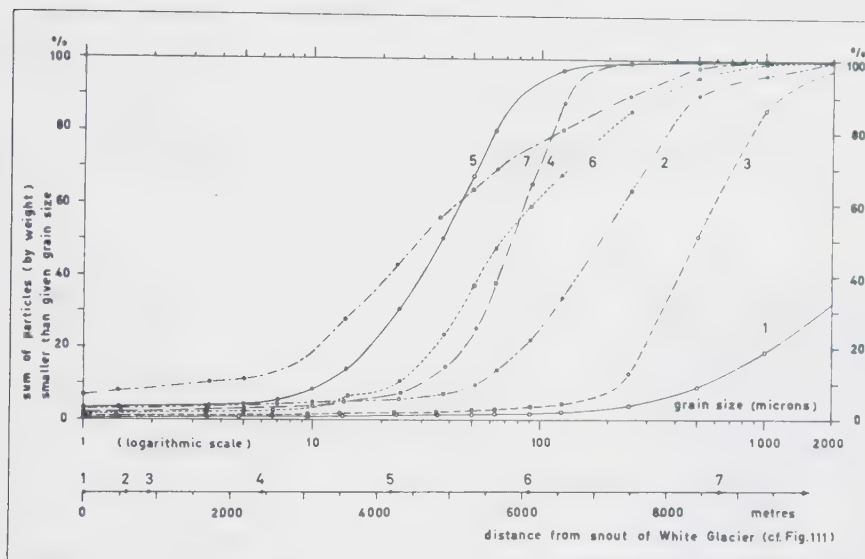


Fig. 117a:
Grain-size distribution of sand samples collected along Expedition River bed.

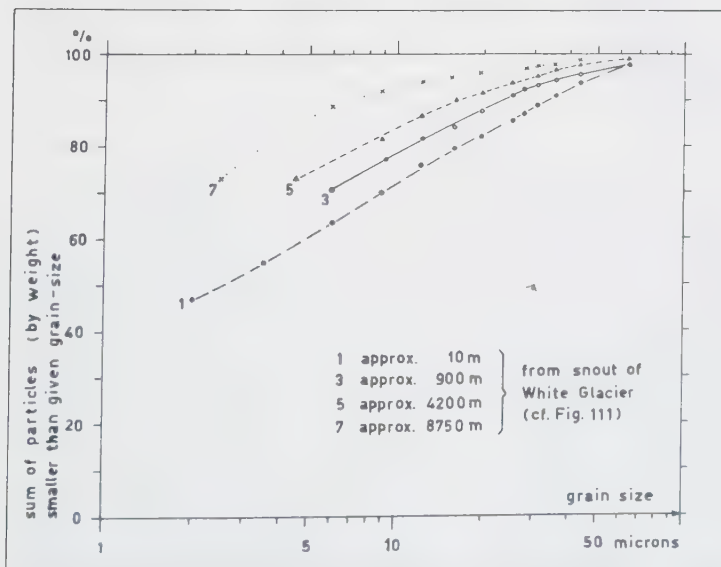


Fig. 117b:
Grain-size distribution of suspended load. Samples collected in Between River - Expedition River at increasing distance from snouts of White and Thompson glaciers, 13 August 1963.

Table 12: Analysis of water samples collected along the Between and Expedition rivers, 13 August 1963⁽¹⁾

Sample No.	1	2	3	4	5	6	7
Approx. distance from White Glacier snout (Thompson Glacier snout) in metres	10	600	900(200)	2400(1700)	4200(3500)	6100(5400)	8700(8000)
Sampling time	0430	0630	0745	1000	1445	1630	1845
Water temperature °C	0.0	0.1	0.15	0.6	2.5	2.7	4.8
SO ₄ mg/1	109.7	-(2)	93.0	51.1	67.5	75.0	80.7
NH ₃ "	0.14	-	0.04	-	0.05	-	0.04
NO ₂ "	< 0.005	-	< 0.005	-	< 0.005	-	< 0.005
NO ₃ "	< 0.25	-	< 0.25	-	< 0.25	-	< 0.25
PO ₄ "	~ 0.01	-	0.02	0.045	≤ 0.02	0.02	~ 0.01
Cl "	0.04	-	30.0	50.0	67.0	158.6	153.0
Ca "	~ 30.0	-	~ 30.0	31.5	42.2	40.0	~ 40.0
Mg "	19.5	-	16.5	11.9	18.9	20.1	14.6
Mn "	0.32	-	< 0.1	-	< 0.1	< 0.1	-
K "	-	-	-	5.1	-	8.0	-
Na "	9.1	-	28.5	15.9	-	56.4	61.8
Fe "	~ 0.02	-	< 0.02	-	~ 0.02	-	~ 0.02
Hardness as CaCO ₃ "							
- Carbonate	10.0	-	32.5	-	50.0	-	45.0
- Total	102.0	-	98.0	-	96.0	-	110.0
Total dissolved substances mg/1 (dry residue)	178	-	211	253	251	465	421
Non-organic % of total	97.8	-	93.4	89.7	84.9	86.7	88.6
Total amount of suspended material mg/1	950(3)	650	3850(3)	6400	5200(3)	4300	3100(3)

(1) Chemical analyses by Kantonales Laboratorium Zürich and Geography Department, University of Zürich

(2) Not analysed

(3) Grain size analyses - cf. Fig. 117b



Fig. 118: T₀-River (2) flowing from T₀ Lake (1) towards Astro Lake (3). Flood water during drainage of Phantom Lake. Lobe of Thompson Glacier in foreground. For situation see Figs 32 and 34. (Oblique air photo, August 1962).

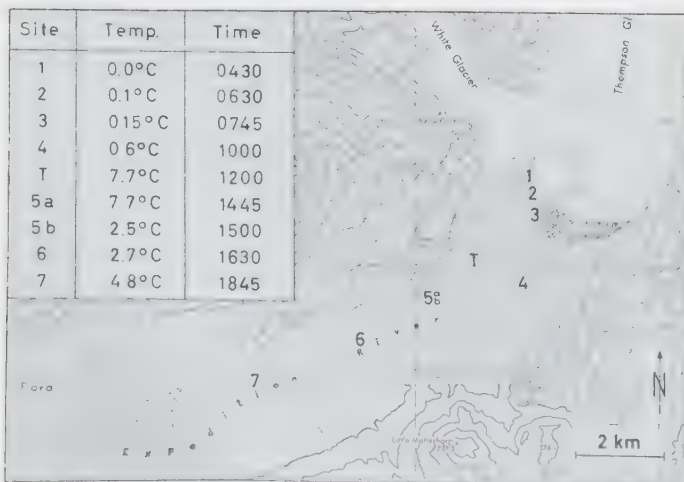


Fig. 119: Water temperatures along Between River - Expedition River, 13 August 1963.

With decreasing slope and increasing distance from the glacier snouts the size of the material deposited by the Between and Expedition rivers is smaller. After a relatively short distance of some 1.5 km only material of sand size was found with no boulders or pebbles. Figure 117a, giving grain-size distribution along the river, shows a decrease of the median size as well as some changes in the sorting of the sands. That the sand sample No. 3 is coarser than No. 2 seems to be due to a strong tributary joining the Between River at this place. The poorer sorting of the sand samples Nos 6 and 7 could result from the addition of new material from the solifluction slopes bordering the outwash plain.

A considerable amount of the load of the Expedition River is carried in suspension. Figure 117b and Table 12 show that down the river the average particle size decreased, whereas the amount per unit of water first increased and then decreased. This could mean that the coarser grains of the suspended material were gradually deposited while new particles were picked up and incorporated.

On 13 August 1963 the water of the Between and Expedition rivers contained relatively large quantities of dissolved matter (Table 12). The solution load increased with increasing distance from the glacier snouts. The large amounts of sulphates and chlorides reflect the strong influence of the nearby evaporite diapirs and the saline springs situated within and at the western border of the Expedition River flood plain (Beschel, 1963a, p. 183-187; Fricker, 1963a, p. 142; Hoen, 1964, p. 37). Sample No. 3, which was collected in 1963 some 200 m from the waterfall that emerged from the Thompson Glacier tongue (Fig. 40), contains abnormally large amounts of chlorides suggesting that there might be some saline springs even further up-valley, perhaps in a part now covered by glacier ice.

The above analyses of the solution load and of the sands it must be remembered, are based on only one or two samples collected at each site, and this in an improvised way. In addition, most of the samples of 13 August 1963 could not be taken along the main branch of the Expedition River but only out of various minor branches on the right side of the outwash plain, and these not simultaneously but one after the other walking down-valley whilst the discharge of the river was slightly increasing.

The pronounced effects on river beds of outflowing lake waters have been demonstrated for the Accident River (Section 1), and particularly for the Between River. The T₀ River valley provides another good example (Figs 32, 118). Only a river with the size and erosive power of that which drains Phantom Lake can create a valley of this size and shape. In the years when there is no emptying of Phantom Lake, T₀ River is but a weak misfit stream.

Ice-dammed lakes, by storing the meltwater of the upper river and glacier reaches and then releasing it in a concentrated mass, help considerably in the shaping of the river channels below. If the same amount of water were to run through the river beds, but distributed over the whole period during which it collects in the lake basins, the geomorphological effect would be very much smaller.

The recurrence interval of such floods is different in every lake-river system depending on the interplay of various factors such as the size of the lake relative to its catchment area, the kind of ice barrier, and the weather. Between Lake has emptied every year since observations began in 1959 except for 1964 when the weather was such that not enough meltwater was produced. So, the recurrence interval of floods caused by the outflow of Between Lake seems to be at present 1 to 2 years, for Crusoe-Baby Lake floods it would be 1 year, whereas for Phantom Lake floods it might be in the order of 2 to 5 years (Chapter B). The frequency of such floods can change relatively quickly following changes in climate or in the height and permeability of the ice barrier, for instance. The speed of the lake emptying and therefore the size and intensity of the flood can vary considerably as demonstrated by the 1961 and 1963 Between Lake drainages. Changes in the recurrence interval and in the magnitude of these floods decisively influence the geomorphological effectiveness of the rivers below ice-dammed lakes.

Wolman and Miller (1960) compare the "work" performed by events of differing magnitude and frequency. They show that it is, in many cases, the events of just moderate magnitude but relative high frequency that performs a large portion of the geomorphic "work". In the Expedition Area, floods produced by rain storms and by the outflow of ice-dammed lakes seem to occur at

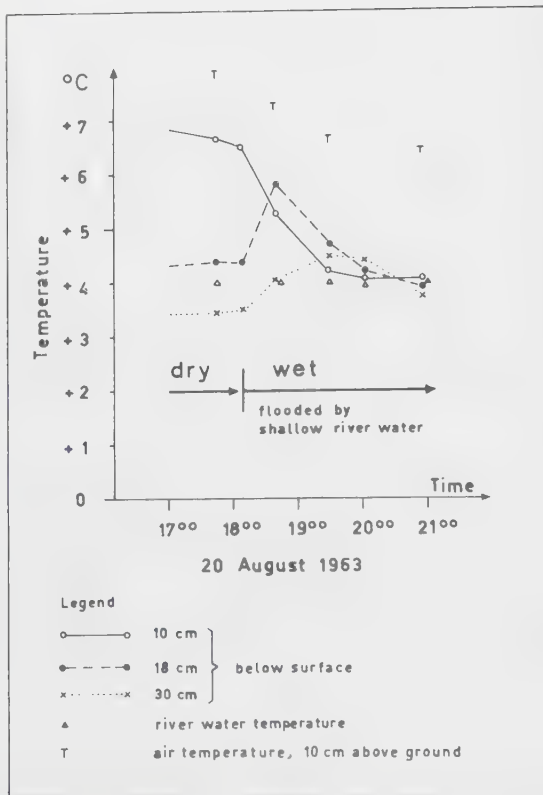


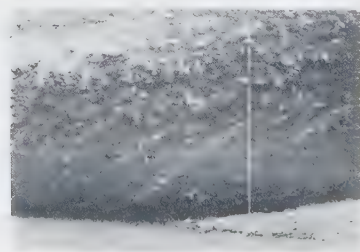
Fig. 120: Near-surface soil temperatures, 20 August 1963, edge of Expedition River gravel outwash plain, before and during submergence by river water. Location of thermistor site T (see Fig. 119).



Figs 121a & b:
Between River, right bank near Point 10:

a) 21 August 1961. Deep under-cut (thermo-erosional niche, some 8 m deep). Dotted line shows bank face 26 June 1961. Man with 10 m pole for scale.

b) Detail of river bank. Horizontally and delta-bedded frozen gravel. 4 m pole for scale.



present relatively frequently; they have been shown to produce moderate to large effects in the river beds. It is therefore suggested that such floods are of considerable importance in the shaping of river channels and outwash plains and, in the long run, of the glacier valley (Chapter E).

4) River Erosion in a Permafrost Area

Axel Heiberg Island is situated in the zone of continuous permafrost and in this zone the thermal effect of river and lake water is particularly important. Some of the forms and changes observed along the Between River and other rivers in the Expedition Area can only be satisfactorily explained by considering this factor.

a) Observed River Water Temperatures

Only a few water temperature measurements in rivers could be obtained but these show clearly, in addition to diurnal and seasonal fluctuations, a warming up of the river water with increasing distance from the glacier snouts, which concurs with observations from other glacierized regions. This effect was observed even towards the end of the season on 13 August 1961, a cold day with almost no sun (Fig. 119). These measurements, however, could not be taken simultaneously at each site; the ones nearer to the glaciers were recorded in the course of the morning, the ones farther away only in the afternoon when the water temperatures are normally a bit higher. The measurements could not always be made in the main branch of the Expedition River, but in several cases had to be taken in secondary branches near the right margin of the outwash plain where the water temperatures are likely to be different. The temperatures measured at the sites 5a and 5b show that there can be considerable variance even in neighbouring branches.

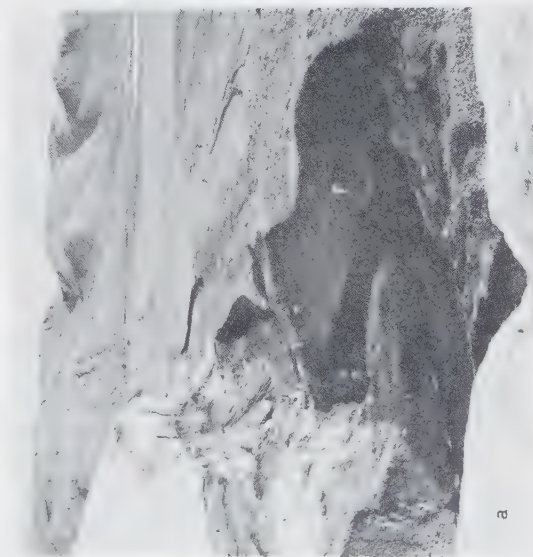
A rough comparison of the few water temperature observations with air temperature, sunshine and sky cover data from the nearby Base Camp station suggests that in the middle of the summer radiation was the main source of heat for the river water, despite the low stand of the sun; towards the end of the melt season air temperature seems to have increasingly influenced water temperatures.

b) The Thawing Effect of Flowing Water

The thermal effect of lake water on the underlying and surrounding ground has been discussed. A river with its constantly new supplies of "warm" water has a similar effect on the thermal régime of the ground over which it flows although this effect is more difficult to assess because of the complicated energy exchange relationships (Legget et al., 1961, p. 963). Only a limited amount of radiation - if any - passes through turbulent river water to reach the ground (Eckel and Reuter, 1950, p. 192-193), it is mainly the exchange of heat between the water and the ground which is of importance. To obtain preliminary information on the thermal influence of flowing water, in August 1963, at three sites on the right margin of the Expedition River outwash plain (Fig. 119, site T), thermistors were inserted into the upper sand and gravel layers to depths of 10, 18 and 30 cm, i.e. all within the active zone which was 50 to 60 cm thick at that time. One of these sites remained dry whereas the other two were flooded by a specially diverted river branch which gave, however, only a very shallow water cover. The weather was abnormally cool and almost without sun during the measurement period. The insertion of the thermistors may have disturbed the permeability and thermal conductivity of the surface layers. Nevertheless, a short temperature sequence, recorded when the site was flooded, is plotted in Figure 120 as it is a confirmation of the results obtained in Between Lake when the basin was submerged by the rising waters. It is the relative behaviour of the temperature in the three layers which is interesting rather than the absolute values. The data show that, in a fairly short time, the uppermost layer took on approximately the temperature of the river water, which was at the time of the diversion cooler than the ground. On the contrary, the temperature of the lower layers at first increased, due in part to the change in the thermal conductivity of the soil on wetting but probably mainly passing through the still relatively warm surface. The temperature differences between the three ground layers were much larger under dry conditions than when flooded by river water. To obtain more precise information on the heat exchange between the river water and the ground and especially on the influence of the water on the permafrost, more elaborate research would be required.

The more effective penetration of thermal energy into ground under or adjacent to water than into bare ground is due to various factors, such as: the higher thermal conductivity of water

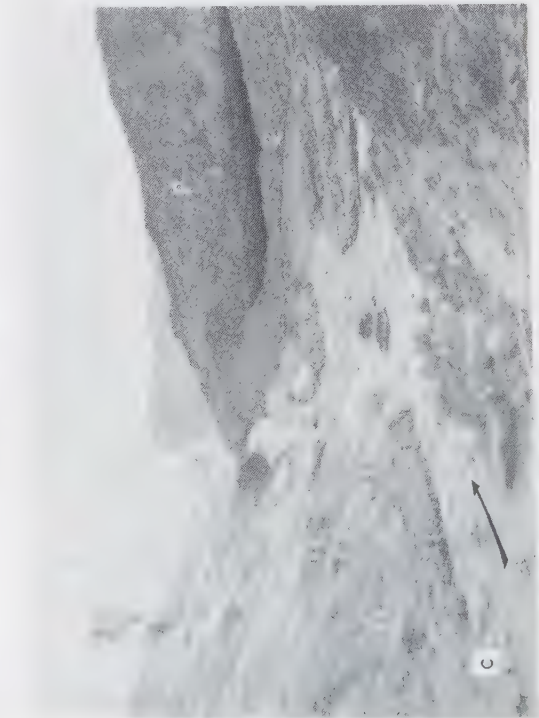
Figs 122a - d: "Prallhang" (P) where Between River cuts through terminal moraine of White Glacier; moraine underlain by old outwash gravel:



a) July 1961:
Between River from
White Glacier snout.
Deep undercut at
bend. Thompson
Glacier snout to left.



b) June 1961:
Between River
dammed up by an
icefall (1) from
Thompson Glacier.



c) August 1964:
As Between Lake did not overflow
in 1964, it was not strong enough
to carry off slumped material, a
ridge of debris developed (2)
sealing off undercut.



d) 29 August 1961:
Detail of the some 35 m deep
thermo-erosional niche.
4 m pole for scale.
(cf. Fig. 123).

and ice, compared with air and dry soil (Sharp, 1942, p. 288-289; Geiger, 1957, cited in Weedfall, 1963, p. 184); the greater specific heat of warm water in comparison with equally warm air and the higher heat storage capacity of water compared with land; the more intensive heat exchange between warm water and a solid than between equally warm air and the same solid (Walker and Arnborg, 1963, p. 10, 13); the higher thermal conductivity and diffusivity of wet sandy and gravelly soil as compared with dry (Geiger, 1957, cited in Weedfall, 1963, p. 184); heat transfer by percolating river water, including the forced convection mentioned by Porkhaev and Sarkisian (1960); and absorption of almost all of the non-reflected solar radiation in a thin surface layer and possible subsequent losses either to the air in the case of free ground or to the turbulent water in the case of flooded ground (Eckel and Reuter, 1950, p. 192-193; Landsberg and Blanch, 1958, cited in Legget et al., 1961, p. 964).

Perennial lakes and rivers moderate the thermal regime of the underlying and adjacent ground with the consequence that the permafrost table is depressed or even missing (Brewer, 1958a; Grigor'ev, 1959b in Legget et al., 1961, p. 963-964; Büdel, 1961, p. 375; Williams, 1963; Johnston and Brown, 1964, p. 163-164). Lakes which freeze to the bottom in winter modify the thermal régime of the permafrost below (Legget et al., 1961, p. 964). This same influence is very probably exerted by ephemeral rivers such as the ones on Axel Heiberg Island although they only flow for some 2 to 3 months in the summer, and are solidly frozen during the greatest part of the year.

The thawing effect of river water is of great geomorphological importance, especially in permafrost areas. The fast deepening and enlarging of meltwater channels - supraglacial, marginal, sub-marginal and pro-glacial - where thawing by flowing water played a decisive rôle, has already been described.

If the river flows over bedrock, thawing can only be effective where the upper layer has been broken up by frost action and ice intersperses the rock fragments as described by Büdel (1961, p. 372) in Spitzbergen. Melting of this ice reduces the cohesion of this layer and lays it open for erosion (Leopold et al., 1964, p. 37-38, 357). In unconsolidated gravel and till which only cohere when frozen, this loosening effect of melting the cementing ice is obvious. At times of higher discharge the river removes the loosened material, and deeper-lying layers of the frozen gravel are exposed to the thermal influence of the water. In this way a more or less uniform lowering and enlarging of the river bed takes place, producing a channel with a relatively broad, flat bottom and steep banks (Fig. 100). Such a cross-section has often been cited as characteristic for present or fossil meltwater channels (von Klebelsberg, 1948, p. 374; Okko, 1956, p. 85, 93-94; Charlesworth, 1957, p. 459; Ives, 1959, p. 58; 1960, p. 54; Derbyshire, 1962, p. 1117) and has also been observed in periglacial river beds (Büdel, 1948, p. 37-38). Channel morphology depends on many varying factors, such as bed material, local topography, and the dynamics and ice temperature of the glacier; however, of overriding importance in all cases is the thawing effect of flowing water.

c) Observations of River Bank Erosion in Frozen Gravel

Most river bank forms discussed here were investigated along the upper reach of the Between River where it cuts through the terminal moraine of the White Glacier and into an old outwash gravel fan (Figs 97-99). The White Glacier moraine and the frozen sands and pebbles that make up the gravel banks of the Between River (Fig. 121b) behave like solid rock. The most striking evidence for the remarkable cohesive strength of these permafrost banks, and for the active lateral erosion of the river, are the deep undercuts and overhanging faces observed in many places along the rivers of the Expedition Area. Based on Russian terminology such undercuts are called "thermo-erosional niches" and the overhangs "cornices" (Gusev, 1952, p. 127-128 in Walker and Arnborg, 1963, p. 9, 14). Due to the more intensive heat exchange between the flowing water and the bank than between the air and the bank the niche is deepened faster than the cornice recedes (Walker and Arnborg, 1963, p. 10, 13).

Along the right bank of the Between River, undercuts of up to 9 m depth, i.e. almost as deep as the bank was high, were measured (Figs 121a, 133b). Thermo-erosional niches of similar size have been reported from Siberia and Alaska (Walker and Arnborg, 1963, p. 10). The deepest undercut observed in the Expedition Area developed in a bend of the Between River, where it cuts through the White Glacier terminal moraine which rests on well-sorted outwash gravel (Figs 122, 123; and Müller, 1963k, p. 170). This thermo-erosional niche in August 1961 measured up to 35 m in depth; some 25 m of frozen gravel and moraine formed the competent cap rock. In 1963 the dimen-

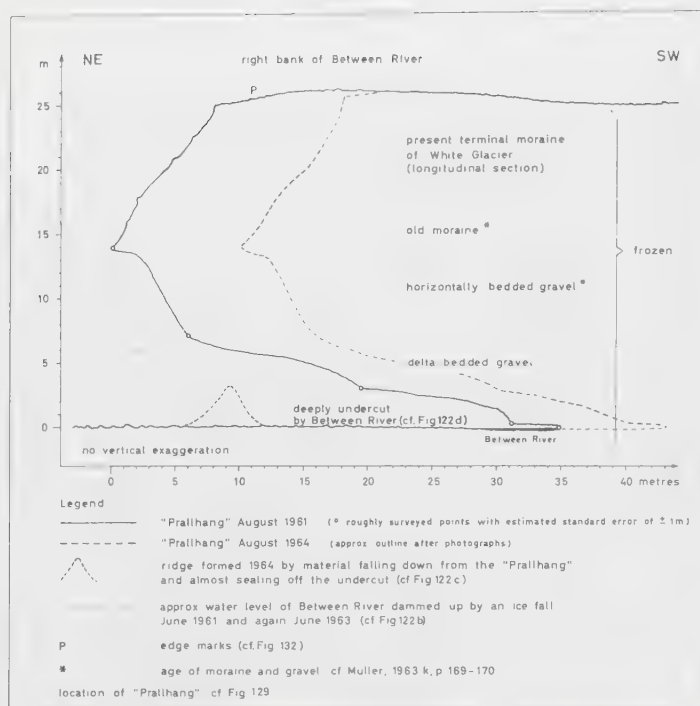
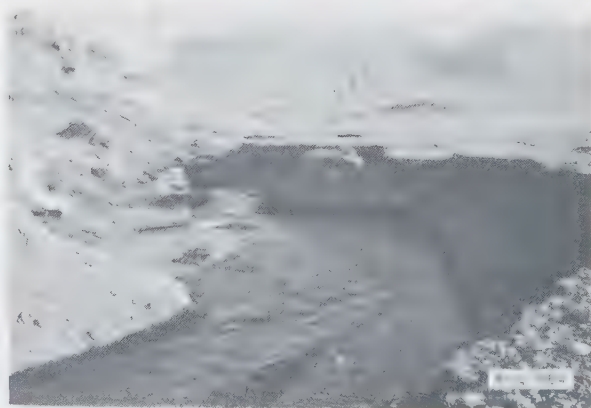


Fig. 123: Approximate section at cut of Between River through the terminal moraine of White Glacier, right concave bank ("Prallhang") with undercut, August 1961 and 1964. (cf. Fig. 122).

Figs 124a - d:
Alterations at right bank (4 m thick frozen gravel) of Between River, near cross-section 24:



a) 28 July 1963: deep undercut in river bend, current still strong enough to remove pebbles, etc., falling from face. b) 3 August 1963: material piles up at foot of bank. c) 12 August 1963: part of undercut river bank has collapsed along an ice wedge (cf. Fig. 125). d) August 1964: slumped gravel block (x) still there as current not strong enough to remove it (no emptying of Between Lake in 1964). Banks are less steep than previous year when river was actively undercutting.



Fig. 125: Frozen river bank (horizontally bedded gravel) collapsed along ice wedge, 12 August 1963.



Fig. 126: Collapse of a permafrost gravel bank in a meander bend of Between River, 10 July 1963, 1300.



Fig. 127: Right bank of Between River at edge points 12 and 13. Poles show total amount of recession of upper edge from 23 July to 9 August 1961. Main current shifted from site. Part of undercut sealed by slumped material. Note steep angle of upper part of bank face.

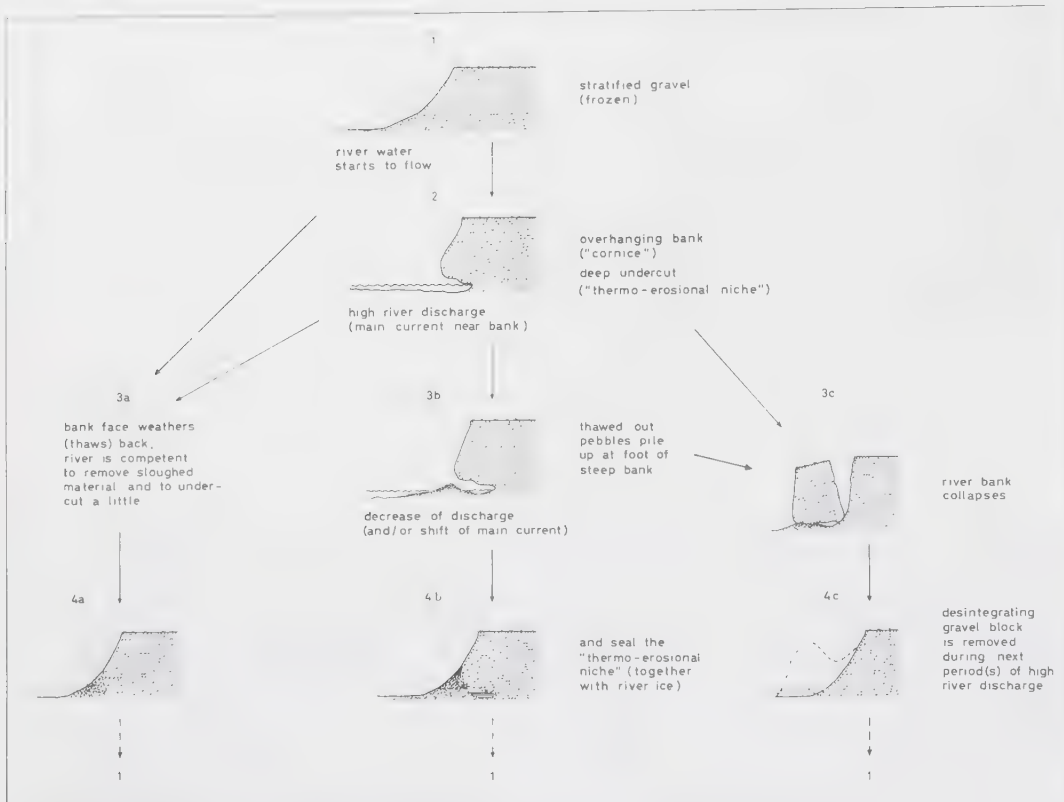


Fig. 128:
River bank recession in frozen gravel, activated by lateral erosion of glacial stream. (Diagrammatic cross-sections based on observations at Between River).

Fig. 129:
River bank recession, summer 1961, selected points along upper edge of right bank of Between River.

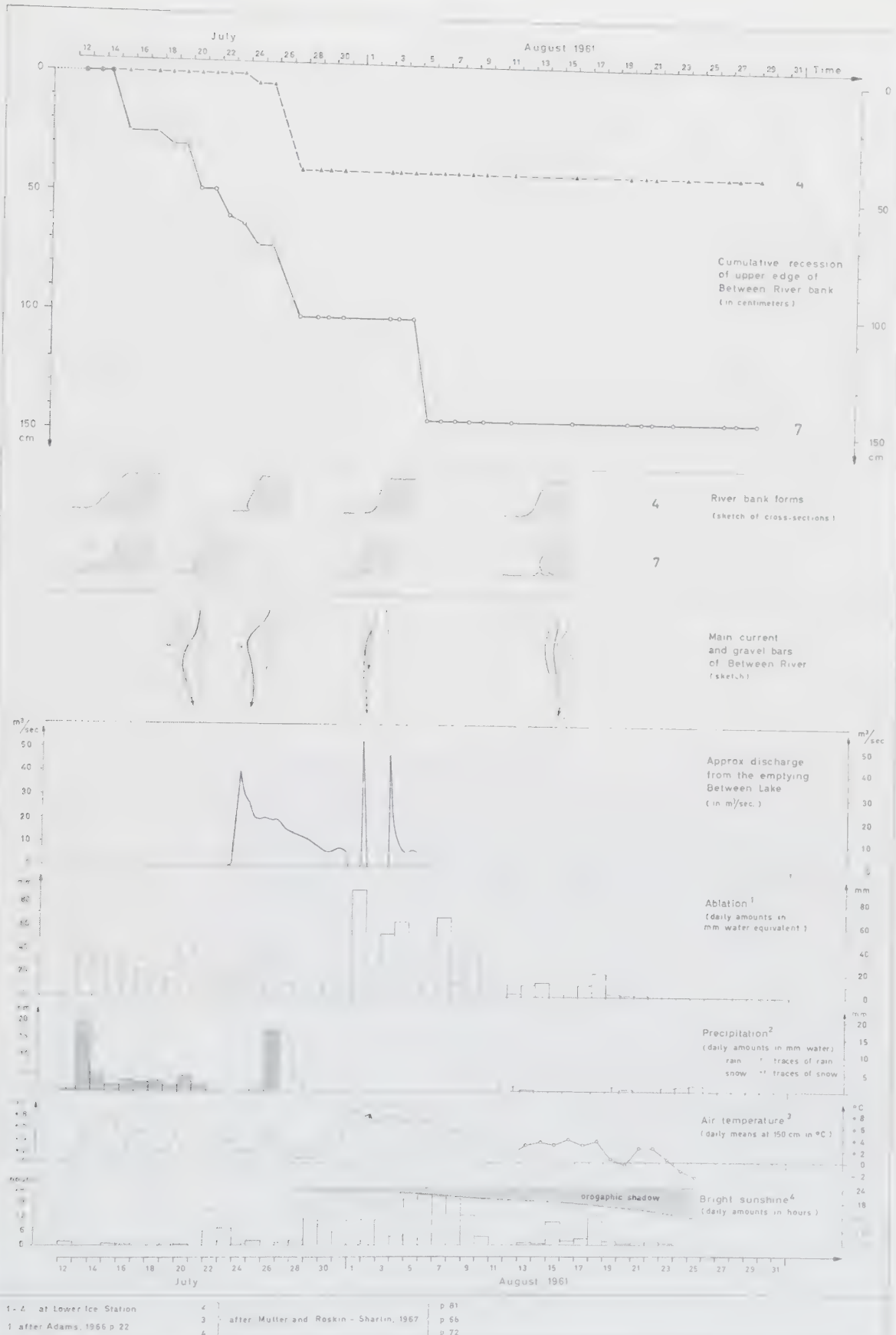


Fig. 130: Recession and forms at two points of right side of Between River and comparison with river course, discharge, ablation values and meteorological factors, summer 1961.

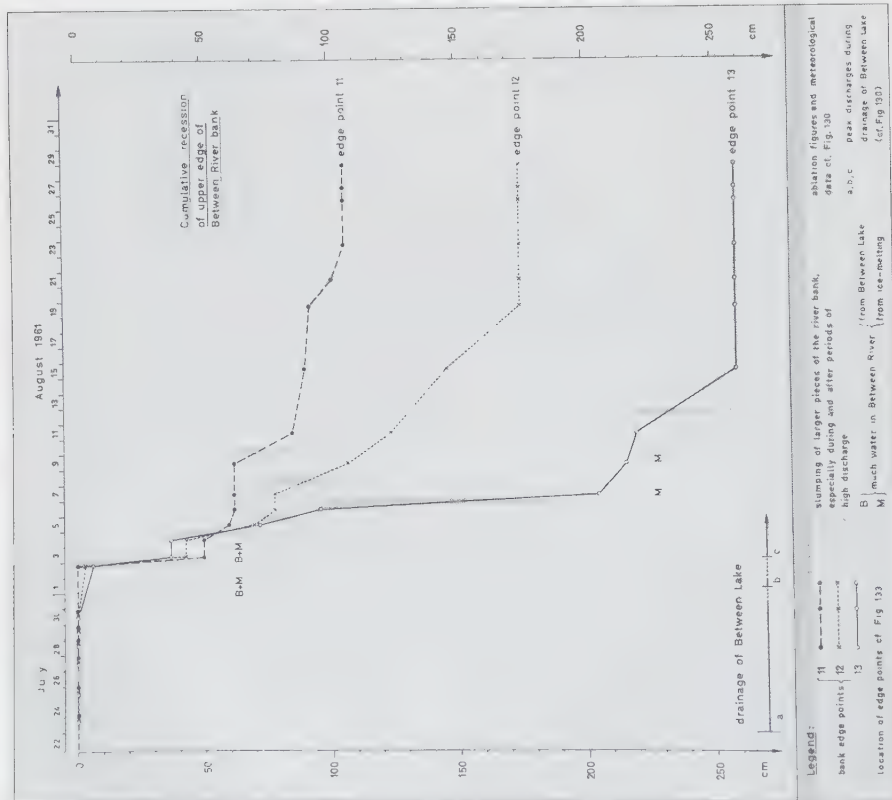
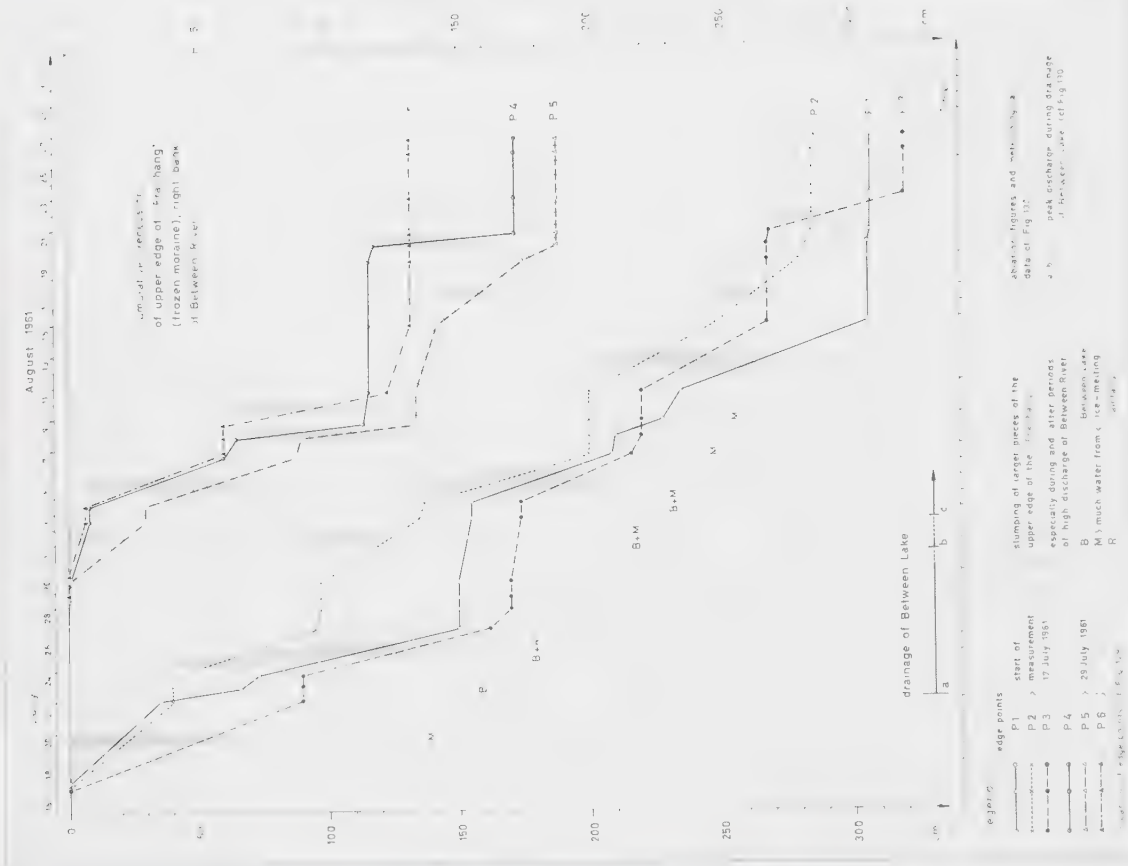


Fig. 131: Recession at points 11-13, upper edge of right bank of Between River, summer 1961.

Fig. 132: Recession of upper edge of "Prallhang", right bank of Between River, summer 1961.



sions of this niche were estimated to be about the same. The erosion of this tremendous undercut by the Between River, the so-called "Prallhang", was the work of several summers. Besides thawing and mechanical erosion by flowing water, the enlarging of this niche was also furthered by the thermal influence of standing water when a lake was formed by ice blocks that fell from the Thompson Glacier cliff (Fig. 122b) damming the Between River.

At low volume, when the main current of the river shifts away from an undercut bank, the thawed-out material is not removed but piles up at the foot, in front of the entrance to the undercut. Sometimes a small part of the river continues to flow under the cornice, but usually fairly soon the thermo-erosional niche is sealed off completely. This development, observed at many places along the Between River and other rivers in the Expedition Area (Fig. 124), is described by Walker and Arnborg in Alaska (1963, p. 13). Even the huge undercut in the bend was almost sealed in the summer of 1964. That year Between Lake did not overflow and ice-ablation was much reduced. The Between River was not strong enough to remove the material that fell down from the face of the undercut moraine, so a ridge of thawed out moraine and gravel debris piled up at the entrance to the thermo-erosional niche, forcing the river to by-pass the undercut (Figs 122c, 123). Some of the sealed undercuts are opened again and occupied by a later flood or because of a shift in the main current, either during the same summer, or only in following years (Walker and Arnborg, 1963, p. 21-22). It was difficult to decide, frequently, whether a thermo-erosional niche was newly cut or just re-opened.

Some of the deeply undercut cornices collapsed (Figs 124c, 126) and then disintegrated slowly (Fig. 124d) or were removed by subsequent flood waters. Figure 125 shows part of the Between River bank that had collapsed along an ice wedge. Walker and Arnborg (1963, p. 14, 15, 23) report that such breaks along ice wedges occur rather frequently in the delta of the Colville River, Alaska. The depth to which undercutting can proceed before the overhanging cornice collapses appears to depend much on the height of the bank, besides the cohesive strength of the frozen bank material. At edge points 12 and 13 (for situation see Fig. 133) where the banks are 2 to 3 m high (Fig. 127) the thermo-erosional niche was never deeper than about 1 m, whereas at edge points 9 and 10 with a bank height of 9 m (Fig. 121), the undercut was some 8 m deep at the end of the summer, and at the 25 m high "Prallhang" the thermo-erosional niche had again a depth of almost 35 m (Fig. 123).

The frozen gravel banks of the Between River had in the upper part average angles of 50° to 70° while the lower part was near vertical or even overhanging. Banks that had recently lost large pieces by slumping had very steep faces. Undercutting and slumping produce much steeper faces in permafrost gravel than in banks of similar but unfrozen material. The slopes of banks not touched by a competent river were usually less steep, and, contrary to the undercut banks, their mean angle of slope declined throughout the summer, due to the thawing and subsequent sloughing of surface layers which accumulated at the foot, at an angle of 30° to 40° , protecting the lower parts from further thawing. Not only the mean angle of slope but also the form of the bank face depends on the extent of river action.

In Figure 128 some of the river bank changes observed along the Between River and other rivers in the Expedition Area are summarized in diagrammatic form.

To measure the rate of recession of permafrost river banks, in the summer 1961, paint lines were drawn parallel to the upper edge of the right bank of the Between River and 16 reference points fixed on these lines with poles and rocks. The distance from these reference lines and points to the bank edge was measured frequently, though at irregular intervals, during the 1961 summer and again once in 1962. In addition, the distance from the cairns of the surveyed cross-sections to the bank edge was checked several times in 1961 and 1963, and less frequently in 1964 and 1965. A series of photographs from marked standpoints and observations on the angle of the bank faces, cracking and the course of the main current give supplementary evidence on bank changes.

The cumulative recession is plotted in Figures 129-132. The slumping of compact gravel pieces and the collapse of whole sections of the bank accounted for most of the recession. At edge point 13, for instance (Figs 127, 131), the total recession of summer 1961, some 2.50 m, occurred within a period of less than two weeks, and almost half of this was due to one slumping on 7 August. Recession of the upper edge did not always give a true picture of the volume changes that occurred on the bank. The upper edge of the right bank (cross-section 21 in Fig. 100) receded little during



Fig. 134: Between River bed (26 June 1961) with thick layers of river ice. First meltwater cutting into river ice. Note slough debris covering some of the river ice.

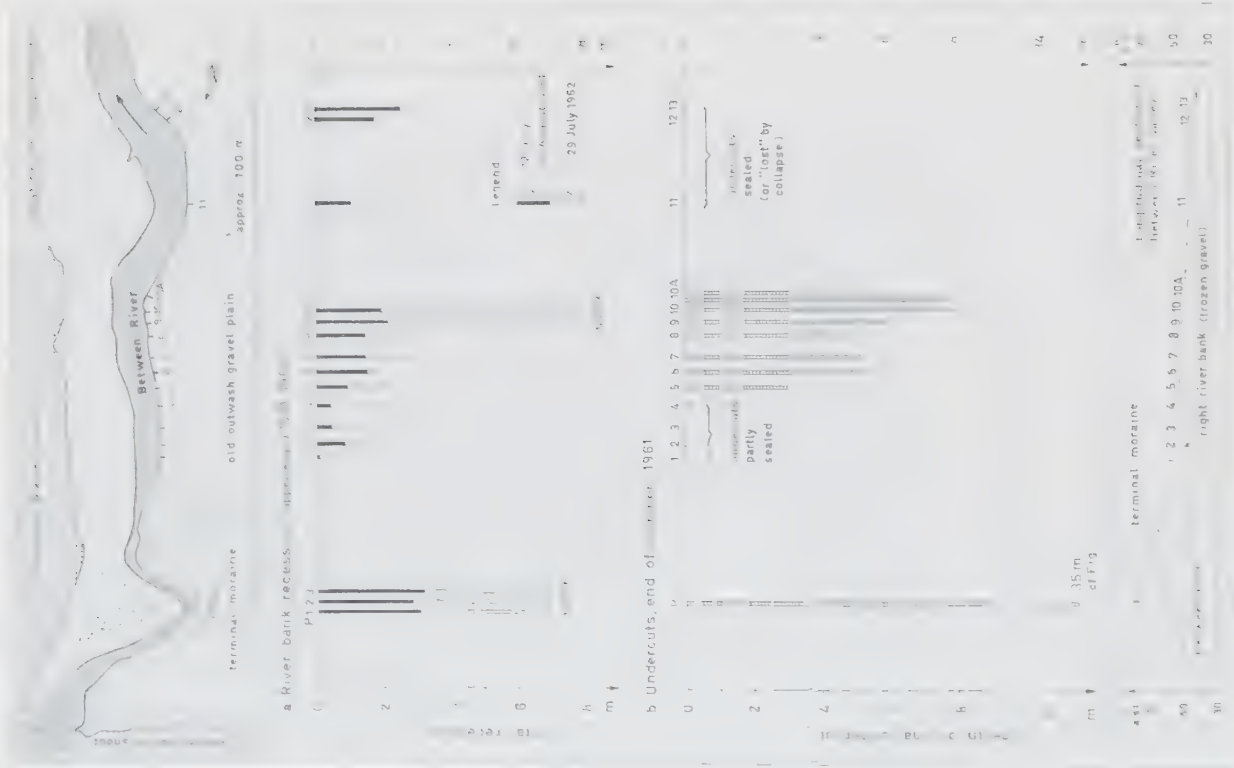


Fig. 133: Total recession of upper edge of right bank of Between River, selected points, 12 July - 29 August 1961 - 29 July 1962, and depth of undercuts (thermo-erosional niche) at end of summer 1961.

the summer 1961, whereas the face changed its form and suffered a substantial loss of material through thawing and sloughing and undercutting at the base.

Heaviest slumping occurred during and after periods of high water, i.e. especially during the drainage of Between Lake, and at times of major rainfall and strong ablation. Somewhat surprisingly the speed of recession of the upper edge of the "Prallhang" was more or less dictated by the river discharge, although the measuring points (Fig. 123) were situated a considerable distance above the thermo-erosional niche.

Besides a possible correlation between river discharge and the rate of edge recession, differences in the rhythm of bank erosion are shown in Figures 129-131. Edge recession during the summer 1961 was clearly influenced by the position of the main current relative to the site and by the location of gravel bars and debris protecting the base, factors which may change from day to day. The changing river topography produced by varying discharge results in differential edge recessing, as was observed along the Between River (Figs 129, 130).

The total amount of recession varied from point to point along the river (Fig. 133a). In the summer 1961, at site 1, the upper edge of the gravel bank receded only a very few centimetres, whereas at sites 9, 10, 12 and 13 some 2 m was lost, and at sites P1, P2, P3, on the "Prallhang" in the terminal moraine, the recession was approximately 3 m. The mean cumulative recession of the edge points 1 to 13 was about 1.20 m. In 1962, total recession at each individual point was different from that of 1961, but for the whole reach the mean recession was of the same magnitude. An exceptionally large retreat occurred at the sites 9 and 10, where the edge receded some 6 m in that year, because of the collapse of a deeply undercut bank. It must be remembered that in these two years the push of the advancing Thompson Glacier forced the river to attack the right bank rather than the left. Nevertheless, in 1961 and 1962, the banks of the Between River seem to have changed as fast as those investigated by Wolman (1959) and Twidale (1964) in temperate regions. A still larger mean annual recession than that recorded along the Between River was observed by Walker and Arnborg (1963, p. 15) in Alaska.

The comparison of edge measurements recorded at the end of August 1961 with photographs from the first part of the summer of 1962 shows that during the 9 to 10 months when no water flowed through the Between River little recession occurred. An occasional collapse occurred during the winter, as reported from other arctic regions (Stefansson, 1961, p. 345, in Walker and Arnborg, 1963, p. 23).

The total amount of recession during the summer of 1961 compared with the depth of the as yet not sealed thermo-erosional niches measured at the end of that summer (Figs 133a, b) indicates that at most sites the recession of the bank edge proceeded more slowly than the deepening of the undercuts, assuming that they were not just re-used but were eroded entirely in the summer of 1961.

It has been stated in older reports from the Arctic (Tarr, 1909, p. 127; Gorbatskii, 1935) and also in a recent one (Vivian, 1965, p. 377) that the hardness of permafrost and the shortness of the run-off period minimizes fluvial erosion. The observations from the Expedition Area show that erosion of permafrost by ephemeral meltwater rivers can be remarkably effective. Rudberg (1963b, p. 146, 148) points to the widespread evidence of intensive fluvial erosion on Axel Heiberg Island, and observations in Spitsbergen, Alaska, and the Russian Arctic confirm this finding (Büdel, 1961, p. 371-73; Walker and Arnborg, 1963). The characteristic peaked discharge of the rivers aided by the freeze-thaw action on the ground is largely responsible for the effectiveness of fluvial erosion in permafrost areas. The thermal preparation of the ground beneath and adjacent to the river takes place when the water temperature is only slightly above 0°C . In 1961 and 1963 the temperature of the Between River water did not exceed 0.5°C .

5) River Ice and Fluvial Erosion

At the beginning of the run-off season in late May or early June (Adams, 1966, p. 59, 60), in 1960, 1961 and 1963 the first river water was observed in many places to flow not on bed material but on river ice and snow, as has been reported also from other arctic areas (Maurie, 1953, p. 703, in Cailleux, 1954, p. 306; Keeler, 1959, p. 95-96; Sim, 1962). The extent and thickness of this river ice varied considerably within the same river bed and from river to river often levelling off



Fig. 135: "Aufeis" covering Expedition River outwash plain, spring 1967. The ice formation originated from a late overflow of Between Lake in 1966. Part of push moraine in foreground.

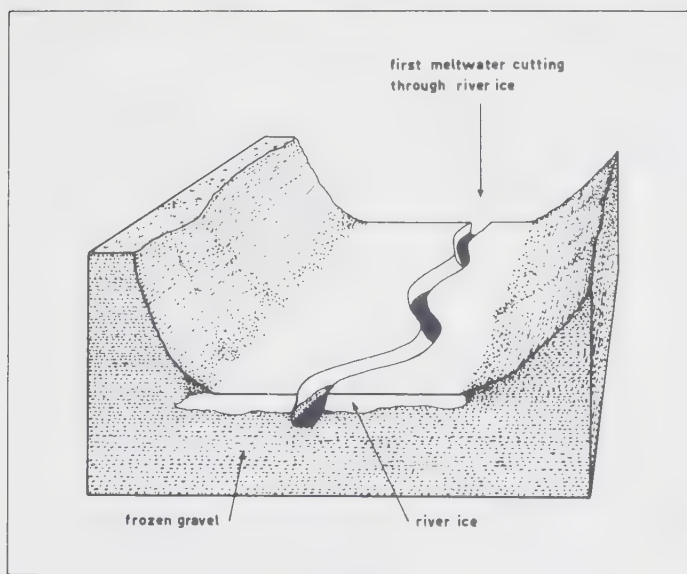


Fig. 136: "Superimposition" of river course through ice on bed (diagrammatic).

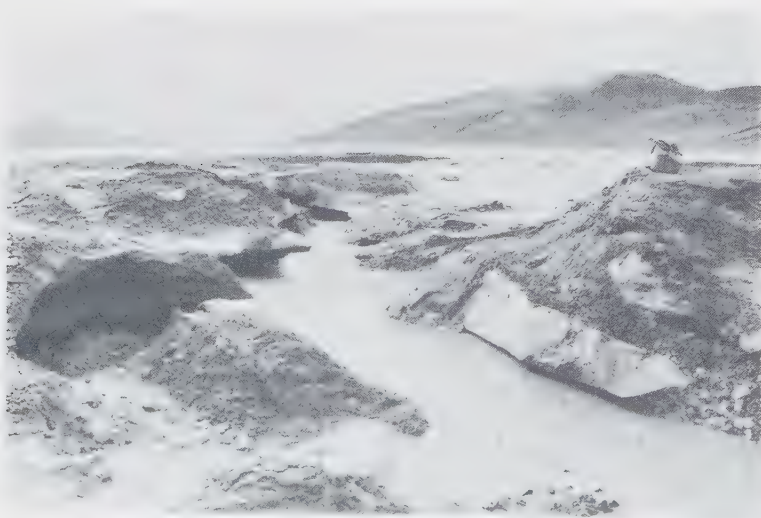


Fig. 137: Perennial ice in the Expedition River outwash plain. Layers of fluvial sediments on top of ice protecting it from melting. (9 July 1961).

irregularities of the bed. In early June 1960 Adams (1966, p. 59, 60) observed some 45 cm of ice at a site in the Ermine River, and even more ice and snow were present in 1961, when similar amounts were found in other parts of the channel, e.g. at cross-section 3 (Fig. 93). In the Between River bed, near cross-section 21 (Fig. 98a), more than 1.20 m of ice were measured in 1961 (Fig. 134).

Such river ice is formed partly at the end of the run-off season, i.e. in late August or early September, and partly at the onset of the run-off season. The formation of such "Aufeis" was observed at various places in the Expedition Area at the end of May and in early June. The first water to collect in the river beds was pools of local meltwater which froze onto existing ice layers. The first running river water flowed only for a short distance then re-froze, especially in flatter sections of the river bed and during the night hours, or it continued to run subnivally. Particularly thick layers of ice developed at places where the last or first meltwater was ponded by ice falls, bank slumpings, or snow drifts. A spectacular Aufeis formation, covering large areas of the Expedition River outwash plain with an ice layer more than one metre thick (Fig. 135), was observed in spring 1967 (personal communication F. Müller). It seems to have originated from a late out-flow of Between Lake in 1966.

During the first days or even weeks of the run-off period large sections of the river beds are protected by ice layers, also by ice falls and drifted snow, from erosion by the flowing water, as observed at various places in the Expedition Area and elsewhere by Sim (1962). Increased local erosion, however, occurs where the river water has melted a channel through the protecting ice (Fig. 134) and, in concentration, attacks thermally and mechanically the underlying bed material (Fig. 136).

In 1960, 1961 and 1963, most of the river ice had disappeared by early July except for some small remnants that persisted in shaded marginal positions. Ice which had been covered by river sediments or by material that slumped from the banks remained much longer if the protective cover was thick enough and if the river did not directly attack this section. In the middle of the Expedition River outwash plain, thick layers of river ice covered by up to 60 cm of fluvial sand and fine gravel endured throughout the summer (Fig. 137). These ice layers probably originated in a similar way to the thick Aufeis formation observed in spring 1967. They might remain, protected from melting, for years (Stefansson, 1910).

In temperate and sub-polar regions much work has been done on the formation of surface ice, frazil and anchor ice towards the end of the run-off period and their influence on river flow and erosion. The cross-sectional area is reduced and the water is channelized or even forced to flow through tunnels below the ice cover, consequently the water velocity increases and its distribution changes, and - with "siphon conditions" - increased erosion can take place (Hoyt, 1913, p. 7, 31-32, 40-42; Sedo et al., 1944; Wundt, 1953, p. 78; Melin, 1954; Moore, 1957; Tietze, 1961, p. 125-131). On the other hand, it has also been suggested that the newly formed anchor ice protects the river bed (Sim, 1962). The formation of various kinds of late season river ice has been observed in the Expedition Area. To investigate whether or not erosion is accelerated during the freeze-up a study was made at the end of August 1963 on a completely ice-covered, 100 m long reach of the Ermine River. Water samples, taken at the upper and lower ends of the frozen stretch, showed little difference in suspended material content. If "siphon erosion" had taken place the samples from the lower end of the frozen stretch would have more suspended particles than those from above. Destruction of the ice cover revealed that the river was too low to fill out the whole tunnel cross-section, therefore no real siphon conditions existed. The same situation was seen in other rivers and, though observations ceased before the final freeze-up of all rivers, there was no indication that "siphon erosion" in the fall was effective in 1961 or 1963. Locally and in certain years this might be different, as when, after a first strong freeze-up, a belated warm spell suddenly increases the river discharge. Weather data from Eureka and Isachsen, the nearest weather stations, suggest that this is the exception for Axel Heiberg Island. In the Sub-Arctic and in middle latitudes, where the rivers that freeze over continue to flow throughout the winter months, "siphon erosion" due to the narrowing presence of river ice, is very likely more important than in the high Arctic where the run-off stops entirely in winter (Tietze, 1961, p. 131).

Table 13: Filling time of Between Lake (max. water level approx. 195 m a.s.l.) and comparison with ablation values and weather data, summers 1960 - 1964

	1960 4 July	1961 23 July	1962 3 July	1963 23 July	1964 did not empty
Start of overflow of full Between Lake					
Cumulative net ablation on White Glacier up to start of overflow (or end of ablation period*)-cm water equivalent					
a) Moraine Cross Profile (approx. 865 m a.s.l.)	26	0	~32	0	0*
b) Anniversary Cross Profile (approx. 365 m a.s.l.)	~53	~38	~58	~66	~22*
c) Lower Ice Station (approx. 200 m a.s.l.)	62	63	~67	~91	~80*
Average $\frac{a+b+c}{3}$	~47	~34	~52	~52	~34*
Elevation of equilibrium line on White Glacier - m a.s.l.	1200	915	1310	~1200	~400
Bright sunshine at Lower Ice Station - total number of hours from 20 May to start of overflow	496.1	469.2	-	~550** (284 after 12 July)	-
Cloud-free sky (tenths of sky not covered by clouds) - sum of mean daily values at Lower Ice Station from 1 June to start of overflow	109.7	93.3	~106**	~200** (>100 after 12 July)	-
Air temperature at Lower Ice Station - sum of degrees C above freezing of daily means to start of overflow ("melting-degree day totals")	100.3	97.2	~100**	~118**	-
Precipitation at Lower Ice Station - total amount from 1 June to start of overflow in mm of water:					
Snow	1.0	20.5	-**	~7**	-
Rain	traces	65.4	-	~2	-
Total	1.0	97.2	~12	~9	-
Snowfall during previous winter (including spring)	little	more than 1959/60	little	more than 1961/62	much
Number of days with summer snowfall (snow traces):					
June	2(3)	11(1)	0(0)	2(7)**	-
July	2(1)	4(1)	0(1)	0(0)	-

**Partly extrapolated from Base Camp weather data and/or data from Eureka and Isachsen

References: Adams, 1961, p. 70 (Fig. 3), 1963, p. 50-51 (Figs 1, 2), 1966, p. 23, 26 (Fig. 20);
Andrews, 1964, Appendix A; Havens et al., 1965, Appendix B; Müller, 1963f, p. 61, 1966, p. 140, 143, 146;
Müller and Roskin-Sharlin, 1967, p. 43, 66, 73, 81.

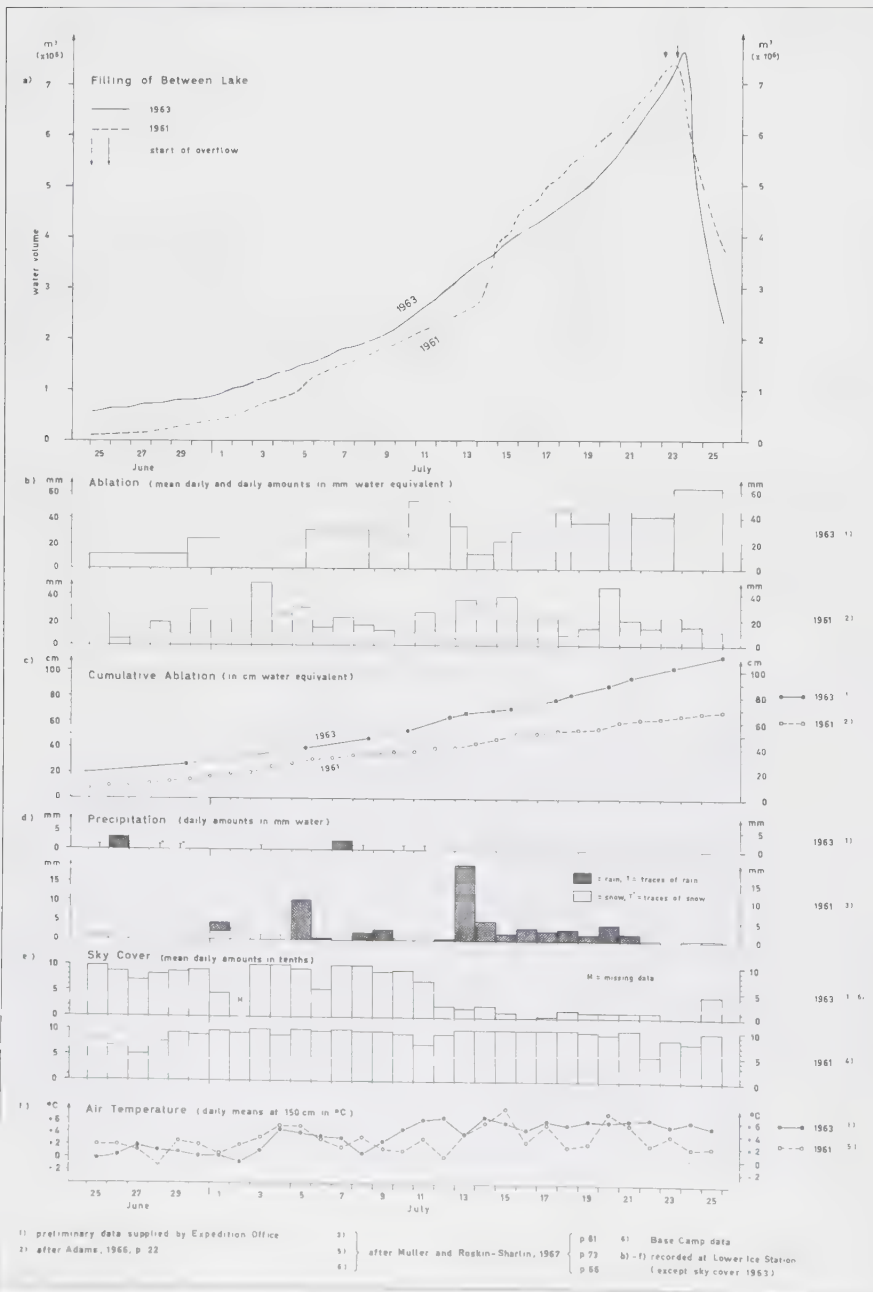


Fig. 138:
Filling of Between
Lake 1961 and 1963,
compared with
ablation and meteor-
ological data of
nearby Lower Ice
Station.

ICE-DAMMED LAKES AND MELTWATER CHANNELS AS CLIMATIC INDICATORS

1) Filling of Present Ice-dammed Lakes related to Weather and Ice Melt

It has been shown that the filling time of any one lake can vary considerably from summer to summer even if it is dammed up to the same level. These variations reflect changes in the weather conditions. In Table 13 weather and ablation data for the period up to the commencement of outflow from Between Lake, in the summers 1960-1964, are assembled. Only sunshine hours and cloud cover observations were available instead of net radiation, and the ablation figures are only approximately representative for the glacierized part of the catchment area of Between Lake, some correspondence with lake filling is apparent. A large part of the inflowing water is glacial meltwater, and a good correlation between net ablation and sunshine and air temperature values has been established (Müller, 1963e, p. 44). So, an ice-dammed lake, such as Between Lake, is a kind of a "run-off totalizer"; from the amount of water in the lake at a particular time, the ablation on the neighbouring glaciers and the weather prior to that date can be inferred.

Precipitation can, however, complicate this relationship (Table 13). In 1961, for instance, when much winter and spring snow remained longer than usual and snow showers occurred even in June and July (Adams, 1966, p. 23; Müller and Roskin-Sharlin, 1967, p. 20-21), the albedo of large parts of the ablation area was higher than in the same period in 1960 and 1962, so that a smaller percentage of the incoming radiation was effective for melt, and consequently less ice was melted particularly in the upper parts of the glacier. If substantial amounts of rain had not compensated for the great deficiency in glacial meltwater supply in 1961, the Between Lake basin would only have been totally filled by the beginning of August. About a quarter of the lake filling was rain water, a much greater proportion than in other summers. In 1963, due to the persistence of winter and spring snow and to June snow showers, the albedo of the White Glacier surface was again relatively high during long periods in June and July, and consequently the feeding of the Between Lake by glacial meltwater was slow, this despite relatively large amounts of cloud-free sky and positive air temperatures. In 1964, after an unusually heavy winter snowfall and a cold, cloudy July (Müller, 1966, p. 143), most of the White Glacier was still covered with snow at the end of the ablation season and Between Lake did not overflow. This means that during the whole summer less snow and ice had melted than "normally" by the middle of July.

More detailed information on ice-melt and weather conditions can be inferred from total filling curves of ice-dammed lakes than from single filling stages. In Figure 138 the filling curves of Between Lake for the summers 1961 and 1963 are juxtaposed. In both years the lake started to empty on 23 July, but differences in the speed of filling during corresponding periods are obvious, reflecting differences in the production of glacial meltwater and in the amount and distribution of precipitation. The importance of the rainfalls in 1961 is evident. The approximate amounts of rain-water and meltwater that contributed to the lake filling can be read out of the curves. If the areas of the catchment and its glacierized part are known, the amounts of rainfall and the ablation values for periods of melt predominance can be estimated.

Ice Cave Lake, Crusoe-Baby Lake and Between Lake usually started to drain in the first half of the ablation season. Their filling curves, therefore, indicate weather and ablation conditions for only the first summer weeks. For information on the rest of the ablation season, curves from lakes that normally drain later are necessary. In 1960 and 1962, Between Lake overflowed almost on the same date, at the beginning of July, which means that up to the end of June the conditions for melt were similar in the two summers. But in July the weather was quite different in the two years; this is reflected by the overflow of Phantom Lake which started almost 4 weeks earlier in 1962 than in 1960 (Table 7).

The deviations from the "long-term" means for the nearby weather stations, Eureka and Isachsen (Müller, 1966, p. 146, Table 2) show that, in 1960 and 1962, June was warmer and less cloudy than "normal", whereas in 1961 and 1963, when Between Lake started to empty some three weeks later, June and July were cooler and cloudier than "normal" (Müller, 1963f, p. 63). In "normal" summers, then, Between Lake will be full about the middle of July if the speed of filling is not influenced by "abnormal" precipitation.

Ice-dammed lakes also show certain other indirect evidence of weather and ablation conditions. On 19 June 1963, there was already a considerable amount of water in the Between Lake

basin and pieces of thin, snow-covered lake ice were floating on it. The thin, broken-up ice cover must have developed after the first meltwater had collected. Prior to 19 June, therefore, there were already days of melt and run-off interrupted by cold spells with minor snowfalls. This interpretation is supported by the weather records at Eureka and Isachsen. The extent and persistence of winter lake ice indicates the prevailing weather. The height of the uppermost supra-glacial and marginal lakes approximates the equilibrium line, as previously discussed.

So, where direct glaciological and meteorological data is lacking supplemental information from ice-dammed lakes can fill the gap.

2) Elevated Shorelines and Abandoned Channels

Field observations, reconnaissance flights and aerial photographs revealed the existence of various sub-recent and relic forms and deposits on Axel Heiberg Island (Robitaille, 1961, p. 174-175; Beschel, 1961a, p. 187-198; 1963b, p. 213-214; Hegg, 1961, p. 201-208; 1963, p. 217-219; Boesch, 1963, p. 165-166; Müller, 1963b, p. 108; 1966, p. 153; Rudberg, 1963b, p. 143). No systematic investigation of such forms was undertaken, however, some examples are briefly discussed, mainly to show that some of the findings from present-day lakes and channels can be used to better understand fossil evidence, and that, on the other hand, fossil evidence can help to explain certain observations on present-day lakes and rivers.

Elevated shorelines, lake sediments, perched deltas and abandoned drainage channels from former or present ice-dammed lakes are clearly of importance to the understanding of glacial change. In cold glacier regions such features indicate the thickness of the ice barriers better than in regions with temperate glaciers (Weidick, 1963, p. 117), as in the former most ice-dammed lakes fill until they overflow and do not empty sub- or englacially. Nevertheless, even in cold glacier areas, shorelines do not necessarily represent former maximum lake levels, as has already been shown.

Former higher lake levels and recently abandoned drainage channels were studied from aerial photographs supplemented by field observations. Although the evidence is not conclusive, it varies from glacier to glacier and between areas within one glacier. The White Glacier has been thinning considerably over the whole length of its ablation area. At the right margin near the outlet of Ice Cave Lake, for example, a lowering of some 3 m in 2 years was registered (Fig. 29) which agrees with measurements of surface change by Müller (1963g, p. 72-75). The tongue area of the Crusoe Glacier, on the other hand, seems to have remained stable or even increased in thickness, as indicated by the various maximum water levels of Crusoe-Baby Lake; but in the upper regions of the ablation area, the Crusoe Glacier may have been thinning recently. The Thompson Glacier, despite its advancing snout, is thinning in the central part of the ablation area (Müller, 1963b, p. 108). Many of the glaciers tributary to the ablation area of the Thompson Glacier show recent thinning. The existence of several large ice-dammed lakes in tributary glacier valleys in different stages of formation is evidence of marked, but differential, thinning and recession of tributary glaciers some yet unknown time ago. This is corroborated by the presence of abandoned meltwater channels and old lake sediments near Port Hendrigan (Fig. 32). The large channels 1 and 2 are good examples of structurally and petrographically controlled fossil river beds. The intakes of these channels lie some 90 m and 53 m, respectively, above the present Phantom Lake water level. The Transit and Phantom glaciers, which once filled the Phantom Lake valley and joined the Thompson Glacier, were so thick that their meltwaters could erode these channels, or, alternatively, the Phantom Lake was once dammed up to this height by a correspondingly thicker Thompson Glacier and the lake overflow water cut the two channels. Lake sediments, where the old meltwater channels join the valley occupied by the T_0 River and an arm of Astro Lake, indicate a former Astro Lake some 20 to 40 m higher than at present dammed by a thicker Astro Glacier when the above mentioned channels were still in use. There is possibly a buried older T_0 channel (Boesch, 1963, p. 166), and abandoned higher sections of the present, partly epigenetic, T_0 River bed.

In Alaska there seem to be relatively more lakes in tributary glacier valleys than in the Expedition Area, which could mean that the tributary glaciers in Alaska have been retreating for a longer time than the glaciers tributary to the Thompson Glacier (Stone, 1955, p. 80).

It has been demonstrated previously that marginal rivers disappear below the glacier margin and sub-marginal and supraglacial rivers leave the glacier at the same place year after year.

The repeated erosional impact will result in cuts into or through moraines and sometimes into bedrock. A good example is the conspicuous gap in the right lateral moraine of the White Glacier near Ermine Lake and the 10 to 15 m deep channel cut into sandstone bedrock (Fig. 139). In 1960, the river, after a sub-marginal stretch of several hundred metres, left the glacier edge just above this cut as it did also in 1961 and 1963. Because of considerable thinning of the glacier, the water now cut a series of minor marginal furrows and benches into the inner flank of the lateral moraine, and entered Ermine Lake first at D and then at E. The moraine cut and the bedrock channel B seem to have been used for a longer time than the minor furrows, suggesting that the White Glacier stagnated for a considerable time. The complex fan at the lower end of channel B indicates that this channel may have been used in more than just one period. Today the channels A-D and their fans and deltas are abandoned.

Another gap through a lateral moraine and various cuts eroded by meltwater rivers are shown in Figure 140. In the summer of 1960 and in following years, the supra- and englacial meltwater from the White Glacier together with a small marginal river flowed through this gap, actively deepening and enlarging it, and into the deeper parts of the Between Lake basin. The meltwater must have left the glacier margin at about the same place for years, even at the time when the edge of the White Glacier was in close contact with the lateral moraine, as evidenced by the size and form of the gap and the higher lying, now abandoned, river cuts (d₁, d₂, d₃). Not only the strong influent D seems to have used approximately the same bed year after year, but also the smaller supraglacial rivers E, F, G, as shown by the cuts (e, f, g), in the crest of the lateral moraine.

Changes in the local base level of the influents into Between Lake, due to the fluctuations of the lake level affect the capacity of the rivers to erode or accumulate. This explains the presence of sandy deposits within erosional cuts.

In the area bordering one of the eastern outlet glaciers of the McGill Ice Cap (Fig. 141), a series of former and present drainage channels, the largest ones being up to 100 m deep, gives evidence of the erosional work of glacial rivers and indicates considerable glacial change. The information and interpretation is taken from the aerial photographs of August 1959, as the area could not be visited.

The larger channels or channel systems, such as 1 to 7, running approximately in a westerly direction, indicate various former positions of the glacier margin. They were probably formed by marginal and sub-marginal meltwaters, eroding year after year along the glacier tongue when it was more or less stagnant or changing only very slowly, as observed at the margins of the present White, Crusoe and Thompson glaciers. Some of the large channels have at times been used simultaneously; as with channels 2 and 3 (in 1959). These two also demonstrate that water flows through them long after the glacier margin has retreated from their lower reaches. The drainage of ice-dammed lakes has also contributed to the erosion of these channels; such a lake existed at the margin of the glacier in August 1959.

The smaller channels, often tributary to the larger ones and sometimes hanging, testifying to a smaller water flow and earlier abandonment, are of less clear genesis. Many of these small channels seem to be approximately "in line" with present supraglacial rivers, and may have been cut by their waters; supporting the assumption of the constancy of supraglacial river courses. Channel 1b would have been eroded by the former river B when the glacier tongue reached several kilometres further down-valley. Similarly, channel 2b was formed when the glacier margin withdrew from the marginal channel 2. In 1959, channel 2b appeared to be already abandoned, and if the recession of the glacier margin from channel 3 continues, a new small channel 3b, tributary to channel 3, will be eroded by the supraglacial river B re-occupying approximately the same bed each summer.

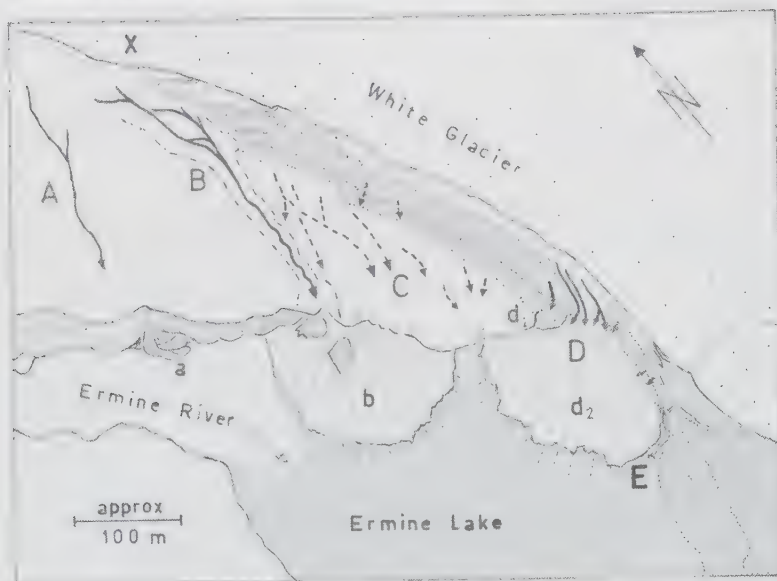
No sign of channels cut by subglacial rivers has been found in this region suggesting that in the past, as at present, the main body of the glacier ice was at a temperature below the pressure-melting point, prohibiting subglacial drainage.

It has been observed that on less inclined slopes more distinct channels are eroded than on steeper slopes where the marginal and sub-marginal rivers often only manage to remove the surface material or to cut a bench. The variation in channel morphology with slope gradient is clearly depicted in Figure 141 and again in Figure 142, showing the slope above the right margin of Crusoe



Aerial photograph, 1954, showing the distribution of meltwater channels near Ermine Lake.

The photograph, 1954, showing the distribution of meltwater channels near Ermine Lake.



Location map.

Legend

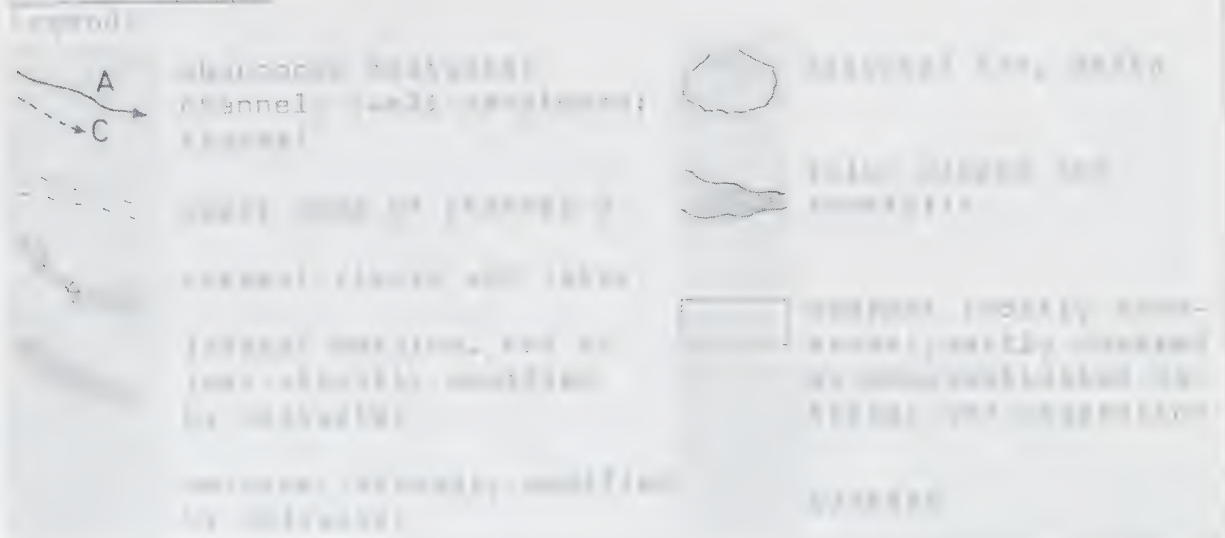


Fig. 139: Abandoned meltwater channels near Ermine Lake.



Fig. 141: Present and former meltwater channels at edge of eastern outlet glacier of McGill Ice Cap.

Glacier. It is evident also that where softer sedimentary rocks form the surface, the meltwater channels are better developed than in the surrounding areas with more resistant igneous rocks.

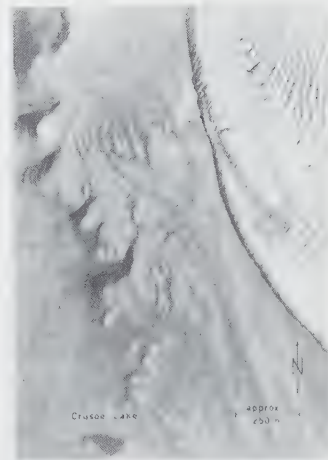
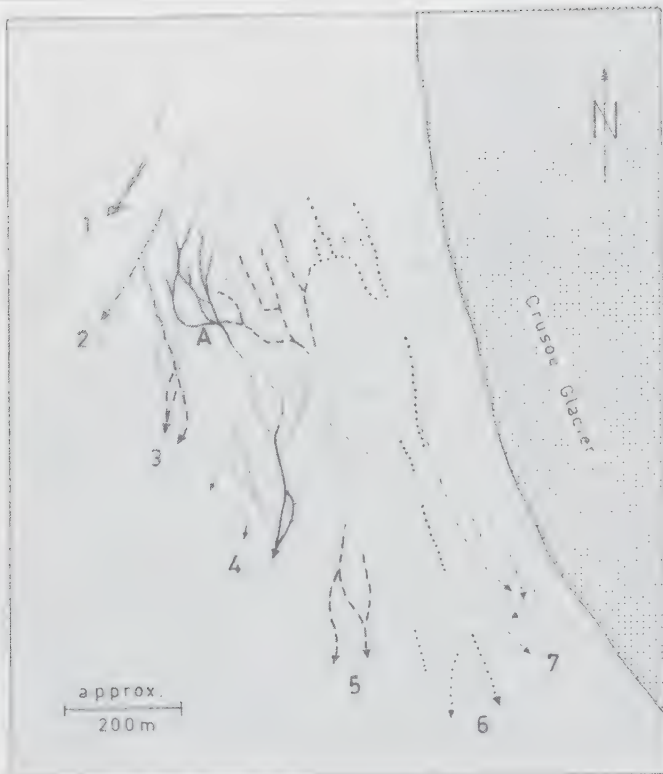
The highest channels in this area (Fig. 142 - cross-section) indicate that the glacier was once at least 125 m thicker than today, and that some of the meltwater flowed over the present water divide into the basin where the Crusoe Lakes are now situated.

Course, gradient, and morphology of these channel systems suggest that they were eroded in a manner similar to the present channels observed along the margins of the Crusoe and other glaciers, that is by marginal and sub-marginal meltwater. So the former Crusoe Glacier was also of cold-ice type. Channel A and some other shorter channels which with the gradient of the slope and at right angles to the margin-parallel furrows, look, at first sight, like subglacial chutes (Mannerfelt, 1945), which would have entailed more temperate ice conditions. However, it is thought that channel A is not a subglacial chute but developed stretch by stretch and not as one channel. This hypothesis is supported by observations elsewhere in the Expedition Area where present marginal rivers have been seen to turn from a strictly marginal course to a down-gradient one penetrating some metres beneath the cold glacier margin and then again turning to flow parallel to the margin, but as sub-marginal rivers. This seems to happen year after year at approximately the same points, usually before or after spurs at structural or lithological lines, or at the junction of tributary land rivers. The short, steep stretch that connects the marginal with the sub-marginal course could be called a "sub-marginal chute". Short down-slope chutes have been observed to develop where marginal rivers cut through the intervening lateral moraine. Channel A coincides approximately with a lithological borderline and a spur. It is, therefore, inferred that the waters that once occupied the marginal channels cut through a lateral moraine and/or penetrated beneath the glacier margin, eroding a short down-slope chute. As the glacier thinned and lower marginal channels were cut, further small stretches of channel A were formed. This erosion took place when glacier ice temperatures were below the pressure melting point, as at present. In the vicinity of Moltke Glacier, Northern Greenland, similar patterns of channels are shown on a photograph published by Schytt (1956, p. 68-69).

The features remaining after a cold glacier has thinned will roughly reflect the gradient of the former glacier margin, as most of the meltwater ran off marginally and sub-marginally. In the case of the Crusoe Glacier, the average gradient of the old marginal and sub-marginal meltwater channels was calculated to be about 1:13, which corresponds approximately with the average gradient of the middle section of the present Crusoe Glacier.

It has been suggested by Mannerfelt (1945, p. 102, 107-108, 225; 1949, p. 196) that many of the fossil channels he investigated in northern Scandinavia reflected the yearly thinning of the ice and were cut in the course of only one summer. But this has been questioned by observations from various areas (Holdar, 1957, p. 369-381; Ives, 1960, p. 53; Sissons, 1961, p. 19; Ives and Kirby, 1964, p. 918) and in a later publication, Mannerfelt (1960, p. 110) considers other than annual origins. It is felt that most of the channels in Figures 141 and 142 are not annual features, as this would entail much too large a yearly glacier mass change. It has been shown that on present glacier margins parts or whole channel systems are re-used year after year and this was certainly also the case when old channels were cut. Small furrows such as those in Figure 142 could have been formed under certain circumstances, as by flood waters, in one summer, but normally they seem to be the work of several summers. The different sizes of the various channels suggest that some have been used much longer than others. The possibility that the Crusoe Glacier re-advanced after periods of recession, and that its meltwaters then used again abandoned channels and cut new furrows among the old ones, must also be considered.

It has been shown in this study that in the Expedition Area the erosional work of marginal and proglacial rivers is impressive, and that it was equally so for former meltwater rivers, erosional traces of which are important evidence of glacier fluctuations, previous drainage patterns, and former glacier ice temperatures. The value of such erosional traces in places where depositional evidence is missing in Alaska, was already recognized by Tarr (1908, p. 102). Recently Boesch (1963, p. 164-165) emphasized that - contrary to temperate glacier regions - in areas with cold glaciers erosional traces indicating former glacier stands are more prevalent and relatively more important than depositional features, as these more easily deteriorate through solifluction. Büdel (1960, p. 80-81) pointed to the destruction of accumulation forms in the Arctic by periglacial mass-movement processes. Nevertheless, depositional evidence was also of great value in the Expedition

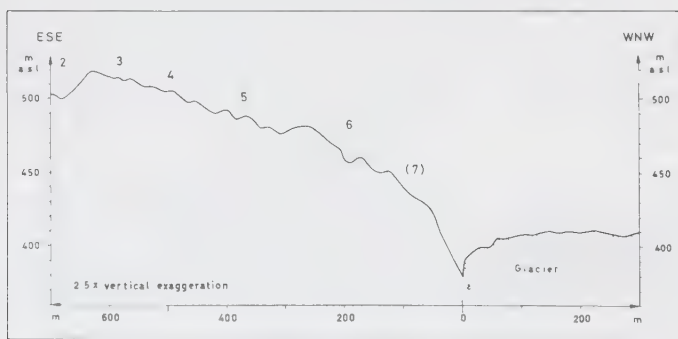


Vertical aerial photograph (part of Nr.4B-55, 2 Aug.1960, RCAF)

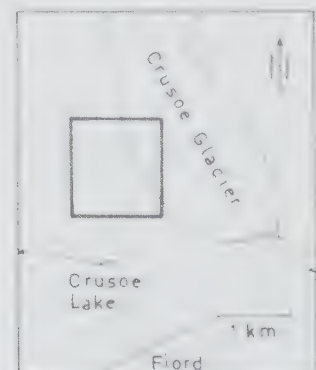
Interpretation

Legend:

- | | | |
|-----------------|--------------------------|--|
| → 1 | higher ones | outcrops of sediments |
| ---→ 2 | systems (probably older) | (mostly shale, Christopher Formation ?), |
| ...→ 3 | of former | surrounded by igneous |
| —→ 4 | meltwater | rocks (gabbro, diabase), |
| - - -→ 5 | channels | cf. Fricker, 1962, Prelim. Geol. Map |
| ...→ 6 | (+ benches) lower ones | |
| - · - · -→ 7 | (probably younger) | |
| — cross-section | | glacier |



Cross-section



Location map

Fig. 142: Former meltwater channels, righthand margin of Crusoe Glacier.

Area (Beschel, 1961a, p. 187-194, Hegg, 1961, p. 201-208 and Rudberg, 1963a, p. 215), especially as within some deposits material suitable for C-14 age determination was found (Müller, 1963k, p. 169-171).

If these observations on ice-dammed lakes and glacial meltwater rivers are applied to the study of Pleistocene glacial evidence in lower latitudes, the limited, localized nature of this present investigation must be born in mind. The environment of the Pleistocene glaciers in lower latitudes, despite certain similarities, was in many respects different from that existing in the Arctic today (Büdel, 1959, p. 301-303; Sater, 1963, p. 92), largely as a result of the latitudinal factor. The most effective variants being the amount and the yearly and daily distribution of solar radiation and of precipitation. This will, of course, induce differences in the hydrological regimen and in the depth of the active layer. Nevertheless, results obtained in an arctic region with cold glaciers, such as Axel Heiberg Island, may help to better understand certain Pleistocene glacial evidence even in lower latitudes. Ahlmann (1953, p. 5) has suggested that most parts of the Pleistocene glaciers were of cold ice type, thus it may be assumed that numerous ice-dammed lakes bordered the glaciers, that meltwater drained mainly supraglacially, marginally and sub-marginally, and that such meltwater rivers played an important geomorphological rôle.

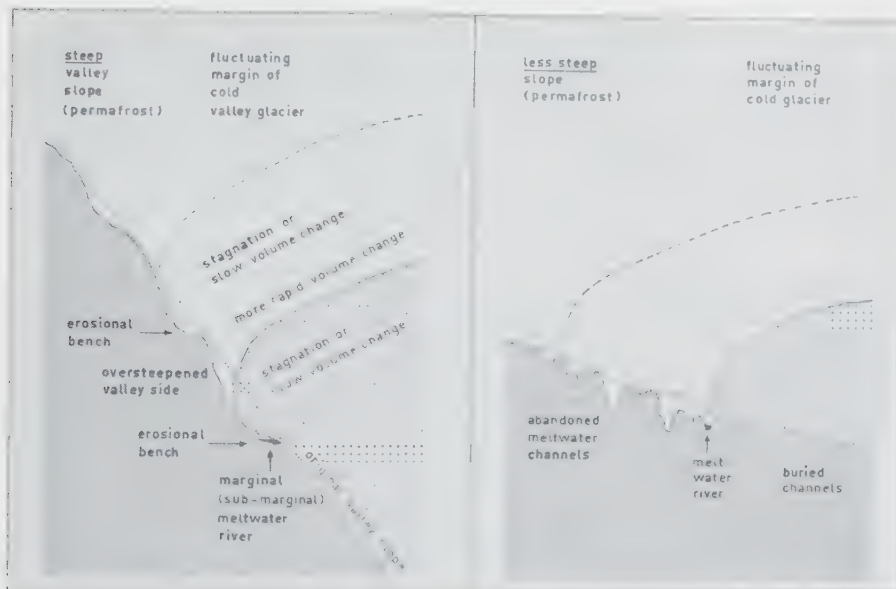


Fig. 143:
Erosion of valley
slope by marginal
and sub-marginal
meltwater rivers
along a cold glacier
(diagrammatic
cross-sections).



Fig. 144:
Snout of White
Glacier with termi-
nal moraine (note
cuts). Fossil
benches (x) on
slope of White
Glacier Hill indi-
cating former posi-
tions of glacier
margin. (24 July
1963).

SUMMARY OF THE GEOMORPHOLOGICAL EFFECTS OF MELTWATER NEAR COLD GLACIERS

The activity of glacial meltwater and its effects differ in regions with cold or temperate glaciers. The fundamental dissimilarities result from the differences in ice and ground temperatures and the hydrological regimen of the meltwater rivers, which are corollaries of climatic conditions.

As previously mentioned, in the Expedition Area practically all of the meltwater runs off supraglacially, marginally and sub-marginally because the main part of the ice bodies are constantly at temperatures below the pressure-melting point. At times minor amounts of meltwater follow locally englacial or even subglacial routes, but, so far, no subglacial drainage of an ice-dammed lake has been observed. Therefore, it is assumed that in the Expedition Area, contrary to temperate glacier regions, there is no, or almost no, meltwater action beneath the main body of the glaciers, but on the other hand, much more along the glacier margins.

Not only the courses followed by the meltwaters within the glacierized valleys of cold and temperate glacier regions differ, but also the hydrological regimen. In the Expedition Area, contrary to temperate areas, there seems to be no winter run-off at all and only a short summer run-off season, "normally" of some 80 days duration, days with 24 hours of sunlight. This concentration of the run-off results in high daily and monthly discharge rates and a correspondingly great geomorphological effect, despite a relatively small yearly run-off total. In addition there is a fairly high frequency of floods occurring after concentrated rainfalls and during the overflow of ice-dammed lakes, which are responsible for the largest changes in marginal and proglacial river beds in the Expedition Area. In the Arctic minor rainfalls can cause powerful flood discharges due to the presence of permafrost and cold ice, and ice-dammed lakes are much more numerous than in temperate glacier areas.

Ice-dammed lakes retain the water from the first part of the run-off season, e.g. Between Lake, or from practically the whole summer or even from more than one summer, e.g. Phantom Lake, and release this stored water concentratedly during relatively short overflow periods. Thus, the already short run-off season is further concentrated to only a few days or weeks of flood water. The effect of this flood discharge is much greater than if the same volume of water flowed through the river bed distributed throughout the whole summer. As most of the lakes empty not at the beginning but rather in the middle or in the second part of the summer at a time when the river beds are free from protective ice, the flood waters can be fully effective. Most of the sediment load of the upper reaches of the river is held back in the lake basin so that the outflowing water is relatively clear, which will affect its erosive capacity in the reach just below the lake (Leopold et al, 1964, p. 454-455). In addition, water that has been stored in an ice-dammed lake has usually been warmed up more than the water in a glacial meltwater river without such a lake, so that the thermal effect of outflowing lake water will be greater.

That the movement of glaciers may be influenced by variations in the supply of meltwater and rainwater has been suggested by various authors (Battle, 1951; Ives and King, 1955, p. 478,482; Haefeli, 1956, p. 12; Meier, 1960, p. 17-18; 1965, p. 762; Glen and Lewis, 1961, p. 1116; Elliston, 1962; Müller, 1963g, p. 68-71; Weertman, 1964; Friess-Greene and Pert, 1965, p. 745-747; Lliboutry, 1965, p. 628; Ostenso et al., 1965, p. 659). Such a correlation is understandable for temperate glaciers where subglacial water acts as a lubricant. Increased summer velocities have also been measured, however, on the surface of the ablation area of the sub-polar type White Glacier (Müller, 1963g, p. 68-71, and personal communication) and on a glacier in eastern Greenland, believed to be also of sub-polar type (Friese-Greene and Pet, 1965, p. 739-747). To explain this phenomenon it is concluded that the base of these glaciers is at melting point. The chemical analysis of some water collected at the end of the summer near the snout of the White Glacier suggests that it had been in contact with evaporites beneath the glacier tongue, which may evidence the existence of subglacial water in the tongue area of the White Glacier. The White Glacier shows the greatest variations between summer and winter velocities in its marginal areas, which may result from the presence of sub-marginal meltwater in the summer.

It has been suggested that large ice-dammed lakes, such as Phantom Lake, prevent by thermal and dynamic means the entry of glacier ice into their basins and that the pressure of these huge water bodies also influences the movement of the damming glaciers (Ward, 1953, p. 247; Glen,

1954, p. 316-318; Charlesworth, 1957, p. 456; Ives, 1958, p. 59; Müller, 1963g, p. 71). What Ahlmann (1937, p. 195-196) postulated for temperate glaciers - that during the outflow of ice-dammed lakes the glacier velocity increases - may be worth investigation on cold glaciers also.

In the past, many differing views have been expressed about the rôle of meltwater in the shaping of the valleys of present and former glaciers. Some have suggested that the work of meltwater is only of importance in the proglacial area and is negligible in the still ice-filled section of the valley, others have emphasized the great erosive and accumulative contribution of meltwater in the latter zone also (Charlesworth, 1957, p. 330-331; Tietze, 1961, p. 132-134; Vivian, 1965, p. 395). In discussing the geomorphological rôle of glacial meltwater the importance of the temperature regimen of the glacier and its surroundings has not always been fully realized. As great parts of the Pleistocene glaciers were probably - at least at certain times - of cold-ice type, the interpretation of Pleistocene glacial evidence would be facilitated by a greater understanding of the erosional and depositional processes and forms in present cold glacier regions.

The investigations in the Expedition Area indicate that in the marginal as well as in the proglacial zone of cold glaciers the effect of the concentratedly running off meltwater can be remarkable and that even in the course of one summer considerable changes occur. It has been shown, too, that the same marginal and sub-marginal channels, or at least most sections of them, are re-used for many years, especially in periods where no appreciable glacial mass change takes place. The erosional impact year after year in the same narrow area along the valley slopes will certainly leave its mark. The type of form produced will depend on the duration of erosion, the steepness of the valley slope, and various other local factors; in some places there may be only a washed hillside, in others benches and channels. For a specific glacier-stand, a much more complete string of erosional and depositional evidence of meltwater action results along the whole ablation area of a cold glacier than is the case with a temperate glacier. So the thickening and thinning of a cold glacier, situated in a valley with relatively steep sides such as the Crusoe, White or Thompson glaciers, is reflected in the "wandering" of the erosion traces up and down the slope. The effect is an over-steepening and cutting back of the valley sides, and the initiation of gully and rill erosion, as suggested by Rudberg (1963b, p. 142), which further steepens the valley sides. Perhaps such marginal water action probably contributed substantially to the erosion of the steep walls of certain troughs and fiords characteristic of some arctic, sub-arctic and even temperate regions, assuming the presence of cold glaciers. Where long periods of glacier stagnation or very slow minor fluctuation are followed by periods of rapid mass change, the valley slopes will develop a step-like formation (Fig. 143). Terraces on the southeast-facing slope of White Glacier Hill, some 100 to 150 m above the present margin of the White Glacier tongue (Fig. 144), are believed to be such benches.

According to Rudberg (1963b, p. 139, 146) and Müller (1963b, p. 99-100) evidence of the erosive and accumulative activity of running water is far greater in the Expedition Area and in other areas of Axel Heiberg Island than would be expected from the available precipitation and temperature data. An estimate in quantitative terms of the importance of meltwater action relative to other geomorphological processes, similar to the calculations by Jäckli (1957) and Rapp (1960) for areas in Switzerland and northern Scandinavia, cannot be given for the Expedition Area at the moment due to scarcity of data. Nevertheless, it is hoped that these investigations help to assess, at least quantitatively, the geomorphological importance of ice-dammed lakes and glacial meltwater rivers in an area with sub-polar type arctic valley glaciers.

EXTENDED ABSTRACT

Lakes dammed by arctic valley glaciers of sub-polar type and marginal and proglacial rivers were studied in a mountainous, almost 60% glacierized, area of some 900 km² in the west-central part of Axel Heiberg Island, Canadian Arctic Archipelago, 79½°N, 90½°W (Figs 1-3). The investigations were part of the long-term programme of the expeditions led by Professor F. Müller, McGill University, Montreal.

As the main portion of the ice bodies in the investigation area is constantly at temperatures below the pressure-melting point, and because of the presence of permafrost, practically all of the meltwater drains on the surface and along the margins of the glaciers. The concentration of ice-dammed lakes in this area is much higher and the average size of the lakes much larger than in a temperate glacierized region. Grouped according to site, the lakes are of the following types: supraglacial, marginal, intritributary river or glacier valley, at the junction of converging glaciers. It is estimated that in all the ice-dammed lakes of the investigation area considerably more water is stored than is melted during one summer on the glaciers of the same area. More than 90% of this water is ponded in lakes lying between 350 and 500 m a.s.l., i.e., in the lower part of the middle zone of the ablation area (Fig. 10). On the average, the small lakes are situated at higher elevations than the large ones; there are also differences in the average size and elevation of lakes belonging to different types (Fig. 11). There is no evidence of any preferred compass orientation for these lakes, but the location of several ice-dammed lakes is dependent on geological conditions.

Four lakes have been studied in detail:

Between Lake (approx. maximum volume 1963: 7 500 000 m³, approx. maximum depth: 35 m), at the junction of the coalescing tongues of the White and Thompson glaciers, empties almost every year by spilling over the damming White Glacier tongue (Fig. 13, Table 4).

Crusoe-Baby Lake (approx. maximum volume 1963: 260 000 m³, approx. maximum depth: 20 m), dammed up by Crusoe Glacier in a small tributary river valley, overflows every year between glacier margin and valley slope (Fig. 21, Table 5).

Ice Cave Lake (approx. maximum volume 1961: 19 000 m³, approx. maximum depth: 12 m), at the right margin of the White Glacier, empties every year either by marginal overflow or through a sub-marginal passage (Fig. 26, Table 6).

Phantom Lake (approx. maximum volume 1963: 500 000 000 m³, approx. maximum depth: 200 m), largest lake in the area, dammed by a lobe of the Thompson Glacier in a tributary glacier valley, a small part of it empties marginally, but not every year (Figs 32, 34, Table 7).

So far, there is no evidence of subglacial drainage of ice-dammed lakes in the investigation area. The lakes drain normally over the damming ice, or along the glacier margin, or over unglacierized ground at a point which is lower than the damming ice. Outlet channels up to 35 m deep were observed, cut progressively by the outflowing water into the damming ice or between glacier margin and valley slope (Figs 17, 23), mainly by melting. The emptying speed and the maximum discharge vary considerably (Figs 18, 25, 30), probably largely controlled by lake water temperatures and the availability of open outflow channel stretches (Figs 19, 74). The drainage curves of lakes that empty supraglacially or marginally seem to differ from those of lakes that drain subglacially. After a rapid increase of discharge up to the maximum, instead of the sudden decrease shown by subglacially draining lakes, supraglacially draining lakes show a gradual decrease (Fig. 39). The modus and speed of the drainage of ice-dammed lakes seem to be largely determined by the type of damming ice, temperate or cold, and by the temperature of the lake water.

Ice-dammed lakes can indicate glacial change and, in addition, they are "run-off totalizers" and indirect "weather recorders". Given certain reference data from previous years, information on the amount of melt on the neighbouring glaciers and on weather conditions that prevailed can be inferred from the filling records of ice-dammed lakes (Fig. 138 and Table 13).

The sedimentation conditions in periodically draining ice-dammed lakes are discussed. The amount and grain-size distribution of sediments deposited in a lake basin vary considerably from place to place (Fig. 55) and at each particular location during the course of one summer, re-

flecting frequent short-term changes in the depositional environment, such as variations in the discharge of the influents and changes in the local base level due to lake level fluctuations (Figs 53, 54). The grain-size distribution of these deposits differs from that of varves, as a winter layer is often missing and as deposition is much more frequently interrupted and disturbed than in a lake with quiet sedimentation. The average amount (dry weight) of summer sedimentation in Between Lake (one filling, one emptying) was found to be about 130 grams per 100 cm², i.e. a mud layer of some 0.5 cm thickness. Dissolved and suspended material present in the Between Lake water was analysed (Tables 8 and 9, and Fig. 48).

Due to the contact with the damming glacier ice and to the presence of icebergs, the water temperature in the investigated lakes was rather low. In summer 1963, at the time of the outflow of Between Lake, a temperature of about 2°C was observed. Phantom Lake with its many icebergs was even colder, close to 0°C (Fig. 59). In Colour Lake, a landlocked lake, higher water temperatures were recorded (Fig. 61). As this lake is probably meromictic with an abnormal temperature stratification, a simple comparison of its water temperatures with those of ice-dammed lakes nearby is not possible. Nevertheless, it seems that the ice-dammed lakes owe their cold-monomictic (or even amictic) character mainly to the contact with ice and only partly to the high latitude. That a relatively cold ice-dammed lake has some influence on the temperature régime of the uppermost layers of the submerged ground and ice is shown by some thermistor measurements (Figs 65-67). The glacier surface submerged under the 2°C water of Between Lake melted faster than the non-submerged ice surface nearby (Fig. 63).

The extent of strictly marginal drainage along cold glaciers is greater than along temperate glaciers, nevertheless the observations showed that the rivers flow sub-marginally at several places (Figs 72, 75, 76, 88, 89). Examples of various kinds of marginal and sub-marginal channels are given. The collapse of the undercut glacier margin confines the marginal river to a sub-marginal course (Figs 23, 24, 73). Sections of a channel may stay open over winter and may be re-used by the meltwaters of subsequent years (Fig. 74). The repeated work of the marginal and sub-marginal rivers will have some erosional effect even on steep valley slopes (cf. Figs 143, 144). Rivers appear to change from marginal to sub-marginal, or vice versa, at much the same place over the years (Figs 139, 140). Supraglacial meltwater was observed to use roughly the same beds year after year and to leave the glacier at the same points (Figs 140, 141). This may result in cuts through the crest or the whole ridge of a lateral or terminal moraine, and into bedrock (Figs 81, 85, 88, 90).

It is suggested that on, and especially along, cold valley glaciers the moraines are comparatively more modified by water than on and along temperate glaciers. The result of this meltwater action can vary greatly, ranging from moderate modifications of morainic material, such as washing, sorting, rounding and re-depositing, to incisions through moraine ridges and complete destruction of whole sections of moraines (Figs 80-84, 88).

A short analysis of four areas with abandoned meltwater channels and with other evidence of meltwater action shows that those channels were once formed in a similar way to the river beds along the present-day glacier margins (Figs 139-142). Such erosional traces are good indicators of former glacier fluctuations and reflect earlier patterns of marginal, sub-marginal and supraglacial meltwater rivers; they also give evidence of former ice temperatures.

Ice-dammed lakes are retention basins, collecting water from part of the run-off season, or from the whole summer or even more than one summer, and releasing it during relatively short drainage periods. Thus, the already short arctic run-off season is concentrated even more to only a few days or weeks of high discharge. The effect of this flood discharge is much greater than if the same volume of water flowed through the river bed distributed over the entire period during which it collected in the lake basin.

Examples are given of the remarkable modifications that have taken place in the beds of rivers and on the outwash plain, especially during periods of flood discharge caused by rain storms or by the drainage of ice-dammed lakes (Figs 93, 98-109, 112-114, and Table 11). To account for the great erosive power of the flowing water and certain river bed and bank forms, e.g. the deep undercuts, "thermo-erosional niches", (Figs 121-123, 133), the thermal effect of the water in an area of continuous permafrost must be considered. Slumping and collapse after undercutting are the most effective modes of recession of frozen gravel banks (Figs 124-126, 128) and occur on the

relatively few days of overflow of ice-dammed lakes or of major rainfalls and strong ablation (Figs 129-132).

The numerous "lake floods", together with flood discharges after rainstorms, are the main agents in shaping the beds of marginal and proglacial rivers and glacier valleys. Large parts of the Pleistocene glaciers consisted - at least for sometime - of cold ice and were bordered by numerous ice-dammed lakes, with similar effects on the landscape.

Eisgestaute Seen und randglaziale Entwässerung.
Untersuchungen auf der Axel Heiberg Insel,
Kanadischer Arktischer Archipel.

ZUSAMMENFASSUNG

Im zentralen westlichen Teil der Axel Heiberg Insel ($79\frac{1}{2}^{\circ}\text{N}$, $90\frac{1}{2}^{\circ}\text{W}$), in einem gebirgigen, zu fast 60% vergletscherten Gebiet von ca. 900 km², wurden im Bereiche arktischer Talgletscher von "subpolarem" Typ eisgestaute Seen und Gletscherbäche untersucht (Fig. 1-3).

Unterkühltes Gletschereis und Dauerfrostböden verhindern weitgehend ein Versickern des Wassers. Praktisch alles Schmelzwasser fliesst deshalb auf der Gletscheroberfläche oder am Rande der Gletscher ab. An zahlreichen Stellen wird es zu Seen aufgestaut, z.B. auf der Gletscheroberfläche, zwischen Talhang und Gletscherrand, zwischen sich vereinigenden Gletschern oder in durch Gletscher abgeriegelten Seitentälern. Es gibt im Untersuchungsgebiet viel mehr und viel grössere eisgestaute Seen als in einem vergleichbaren Gletschergebiet der Schweizer Alpen. Es wird geschätzt, dass in den eisgestauten Seen des Untersuchungsgebietes total mehr Wasser aufgestaut ist, als während eines Sommers auf allen Gletschern des gleichen Gebietes abschmilzt. Mehr als 90% dieses Wassers ist in Seen aufgestaut, die zwischen 350 und 500 m ü.M. liegen, das heisst im unteren Teil der mittleren Zone des Ablationsgebietes (Fig. 10). Die kleineren Seen liegen durchschnittlich höher als die grösseren Seen, auch gibt es Unterschiede in der durchschnittlichen Grösse und Höhenlage der verschiedenen Seentypen (Fig. 11). Die Seen scheinen keine bestimmte Exposition zu bevorzugen, doch ist die Lage einiger eisgestauter Seen durch geologische Gegebenheiten bestimmt.

Vier Seen wurden etwas genauer untersucht:

Der Zwischen See (Between Lake) (max. Wassermenge 1963: ca. 7 500 000 m³, max. Tiefe: ca. 35 m), aufgestaut zwischen den zwei sich vereinigenden Zungen des White-Gletschers und des Thompson-Gletschers; er entleert sich fast jedes Jahr durch Ueberfliessen der stauenden White-Gletscherzunge (Fig. 13 und Tabelle 4).

Der Crusoe-Baby See (max. Wassermenge 1963: ca. 260 000 m³, max. Tiefe: ca. 20 m), aufgestaut durch den Crusoe-Gletscher in einem kleinen Seitenflusstal; er überfließt jedes Jahr zwischen Gletscherrand und Talhang (Fig. 21 und Tabelle 5).

Der Ice Cave See (max. Wassermenge 1961: ca. 19 000 m³, max. Tiefe: ca. 12 m), am rechten Rand des White-Gletschers; entleert sich jedes Jahr entweder durch randliches Ueberfliessen oder durch einen sub-randlichen Tunnel (Fig. 26 und Tabelle 6).

Der Phantom See (max. Wassermenge 1963: ca. 500 000 000 m³, max. Tiefe: ca. 200 m), grösster See im Expeditionsgebiet, aufgestaut durch einen Lappen des Thompson-Gletschers in einem Seitengletschertal; ein kleiner Teil des Sees entleert sich randlich, allerdings nicht jedes Jahr (Fig. 32, 34 und Tabelle 7).

Bis jetzt wurde im Untersuchungsgebiet kein subglaziales Auslaufen von eisgestauten Seen beobachtet. Die Seen entleeren sich normalerweise über das stauende Eis, oder entlang des Gletscherrandes, oder über unvergletschertes, tiefer als das stauende Eis liegendes Gebiet. Während des Auslaufens entwickeln sich bis zu 35 m tiefe Ausflussrinnen, die - hauptsächlich durch Schmelzen - von der Oberfläche her in das stauende Eis oder zwischen Gletscherrand und Talflanke eingetieft werden (Fig. 17, 23). Die Entleerungsgeschwindigkeit und die maximale Ausflussmenge können von Entleerung zu Entleerung sehr verschieden sein (Fig. 18, 25, 30). Wahrscheinlich spielt in diesem Zusammenhang die Wassertemperatur eine entscheidende Rolle, auch ist von Bedeutung, ob von früheren Entleerungen her noch Reste alter Ueberlaufirinnen vorhanden sind, die erneut benutzt werden können (Fig. 19, 74). Die Entleerungskurven von Seen, die über den Gletscher oder zwischen Talhang und Gletscherrand auslaufen und von Seen, die sich subglazial entwässern, scheinen verschieden zu sein. Nach einer schnellen Zunahme der Ausflusswassermenge bis zu einem Maximum nimmt z.B. beim Zwischensee das Ausflussvolumen - im Gegensatz zu subglazial sich entleerenden Seen - nicht plötzlich, sondern nur relativ langsam ab (Fig. 39). Art und Geschwindigkeit des Auslaufens von eisgestauten Seen scheint also u.a. entscheidend davon abzuhängen, ob der stauende

Gletscher "kaltes" oder "temperiertes" Eis aufweist.

Eisgestaute Seen können wertvolle Anzeiger von Gletscherveränderungen sein; auch sind sie riesige "Abflusstotalisatoren" und registrieren indirekt das Wetter. Falls gewisse Bezugsdaten von früheren Jahren her bekannt sind, kann z.B. aus den Füllkurven solcher Seen die ungefähre Abschmelzmenge auf den benachbarten Gletschern berechnet werden, auch sind Rückschlüsse auf das Wetter, das vorherrschte, möglich (Fig. 138 und Tabelle 13).

Die besonderen Sedimentationsbedingungen in periodisch sich entleerenden eisgestauten Seen werden diskutiert. Es wird gezeigt, dass die Korngrößenverteilung von Seesedimenten innerhalb eines Seebeckens stark variieren kann (Fig. 55). Sie kann aber auch an ein und derselben Stelle beträchtlich variieren, was die häufigen kurzfristigen Veränderungen der Sedimentationsbedingungen (wie z.B. Veränderungen in der Wasserführung der Zuflüsse, Veränderungen in der Lage der lokalen Erosionsbasis, verursacht durch Seespiegelschwankungen) widerspiegelt (Fig. 53, 54). Die Korngrößenverteilungen dieser Ablagerungen scheinen verschieden zu sein von jener "klassischer" Warven, da in den periodisch ausfliessenden Seen eine Winterablagerung fehlt, und da der Sedimentationsvorgang viel häufiger unterbrochen und gestört wird als in Seen, die nicht auslaufen.

Im Zwischensee wurde während eines Sommers (d.h. während einer Aufstau- und einer Auslaufperiode) durchschnittlich ca. 130 g (Trockengewicht) Sand und Ton pro 100 cm² abgelagert; das entspricht einer Sedimentschicht von ca. 0.5 cm Dicke. Gelöstes und suspendiertes Material wurde auch analysiert (Tabellen 8 und 9 und Fig. 48).

Wegen des Kontaktes des Wassers mit Eis (stauende Gletscherwand, Eisberge) sind die Wassertemperaturen in eisgestauten Seen ziemlich tief. Im Zwischensee wurde im Sommer 1963 zur Zeit des Ausfliessens eine durchschnittliche Wassertemperatur von 2°C gemessen. Der Phantom See mit seinen vielen Eisbergen war noch kälter, nämlich nahezu 0°C (Fig. 59). Andererseits wurden im Colour See, der im unvergletscherten Gebiet liegt, höhere Temperaturen gemessen (Fig. 61). Allerdings ist dieser See wahrscheinlich meromiktisch, so dass ein Vergleich seiner Wassertemperaturen mit jenen nahegelegener eisgestauter Seen nicht ohne weiteres möglich ist. Trotzdem ist offensichtlich, dass die eisgestauten Seen ihren kalt-monomiktischen (und z.T. sogar amiktischen) Charakter in erster Linie dem Kontakt mit Eis und erst in zweiter Linie der nördlichen Lage verdanken. Dass auch ein relativ kalter eisgestauter See einen Einfluss auf das Temperaturregime der obersten Schichten eines überfluteten Seebeckens und des Eises haben kann, zeigen einige Thermistoren-Messungen (Fig. 65-67). Es konnte auch beobachtet werden, dass eine unter Wasser gesetzte Gletscheroberfläche schneller schmolz als nicht überflutetes Eis (Fig. 63).

Entlang "kalter" Gletscher gibt es über grössere Strecken randliche Entwässerung als entlang "temperierter" Gletscher. Jedoch verschwinden auch in Gebieten mit kalten Gletschern die Gletscherbäche an verschiedenen Stellen unter dem Gletscherrand und fliessen sub-randlich (Fig. 72, 75, 76, 88, 89). Verschiedene Arten von randlichen und sub-randlichen Abflussrinnen werden besprochen. Es wird gezeigt, wie durch das Zusammenstürzen des unterschrittenen Gletscher-randes der randliche Fluss zu einem sub-randlichen wird (Fig. 23, 24, 73). Teile von randlichen und sub-randlichen Kanälen können über die Wintermonate offen bleiben und eventuell in den folgenden Jahren vom Schmelzwasser wieder benutzt werden (Fig. 74). Die Jahr für Jahr sich wiederholende Arbeit der randlichen und sub-randlichen Gletscherbäche hat auch an steilen Talflanken mit der Zeit eine beträchtliche Wirkung (Fig. 143, 144). Es scheinen immer ungefähr die gleichen Stellen zu sein, wo ein randlicher Fluss unter dem Gletscherrand verschwindet und sub-randlich weiterfliesst, oder wo ein Bach nach einem sub-randlichen Verlauf wieder am Gletscherrand erscheint und randlich weiterfliesst (Fig. 139-140). Auch die auf der Gletscheroberfläche abfliessenden Schmelzwässer benutzen jedes Jahr ungefähr die gleichen Rinnen und verlassen den Gletscherrand ungefähr an den gleichen Stellen (Fig. 140, 141). Durchschnittene Moränenwälle oder Kerben im anstehenden Gestein sind das Resultat dieser ständig an der gleichen Stelle ansetzenden erosiven Tätigkeit des Schmelzwassers (Fig. 81, 85, 88, 90).

Die Ober- und Seitenmoränen von "kalten" Talgletschern werden wahrscheinlich stärker vom Wasser bearbeitet als dies bei "temperierten" Gletschern der Fall ist. Das Ergebnis dieser Schmelzwasserarbeit kann sehr verschieden sein, vom blossen Auswaschen, Sortieren, Runden des Moränenmaterials bis zu tiefen Einschnitten in Moränenwälle oder der gänzlichen Zerstörung derselben (Fig. 80-84, 88).

Eine kurze Analyse von vier Gebieten mit fossilen Schmelzwasserrinnen und mit andern

Zeichen früherer Schmelzwassertätigkeit (Fig. 139-142) zeigt, dass diese Rinnen einst in ähnlicher Weise gebildet wurden wie die in dieser Arbeit untersuchten Bachbetten entlang der heutigen Gletscher. Erosive Spuren (wie diese Rinnen) sind gute Anzeiger von früheren Gletscherständen; sie geben auch Hinweise auf den Verlauf früherer Gletscherbäche und auf frühere Eistemperaturen.

Eisgestaute Seen sind Speicherbecken. Während eines Teiles der Schmelzperiode (oder während mehrerer Schmelzperioden) sammeln sie das Wasser eines mehr oder weniger grossen Einzugsgebietes und geben es dann während einer relativ kurzen Auslaufperiode konzentriert ab. Es ist klar, dass das Abflussrégime von Flüssen unterhalb solcher sich periodisch entleerender Seen stark verschieden ist von jenem von Flüssen, die keine solchen Seen in ihrem Lauf haben. Der in der Arktis ohnehin schon auf einen sehr kurzen Sommer beschränkte Schmelzwasserabfluss wird also in Gletscherbächen mit eisgestauten Seen in ihrem Lauf auf einige wenige Tage oder Wochen mit Hochwasserführung konzentriert. Die Wirkung dieser Hochwasser kann besonders am Rande der Gletscher und im Gletschervorfeld sehr gross sein, meist viel grösser, als wenn die gleichen Wassermengen durch die gleichen Flussbetten, jedoch verteilt über die ganze Periode, während der sich Wasser im Seebecken ansammelte, fliessen würden.

An Beispielen werden die bemerkenswerten Veränderungen gezeigt, die in den Betten verschiedener Flüsse und in einer Sanderebene des Expeditionsgebietes beobachtet werden konnten (Fig. 93, 98-109, 112, 114 und Tabelle 11). Die grössten Veränderungen fanden bei Hochwasser, verursacht durch starke Regenfälle, starke Ablation oder See-Entleerungen, statt. Die grosse erosive Wirkung des fliessenden Wassers, wie auch verschiedene spezielle Erosionsformen, z. B. die Kastenform der Schmelzwasserrinnen und die tiefen Unterschneidungen ("thermo-erosional niches" = "Schmelzerosionsnischen", Fig. 121-123, 133), können nur richtig verstanden werden, wenn berücksichtigt wird, dass in diesen Dauerfrostbodengebieten die Schmelzwirkung des Wassers von grosser Bedeutung ist. Das Herunterfallen kompakter Uferstücke und das Abbrechen ganzer, tief unterschrittener Uferpartien verändern Ufer aus gefrorenem Schotter am stärksten (Fig. 124-126, 128). Diese Vorgänge wurden am häufigsten während oder nach Hochwassertagen beobachtet (Fig. 129-132).

Da in der Arktis eisgestaute Seen zahlreich sind, tragen "See-Hochwasser" - neben "Regen-Hochwassern" - wesentlich zur Ausgestaltung der Betten randlicher und proglazialer Flüsse und damit zur Formung des Gletschertales bei. Grosse Teile der eiszeitlichen Gletscher hatten - wenigstens zeitweise - kaltes Eis, und zahlreiche Seen wurden an ihren Rändern aufgestaut und entleerten sich wohl in ganz ähnlicher Weise wie die Seen, die in dieser Arbeit beschrieben wurden. Ihre Hochwasser übten am Gletscherrand und im Gletschervorfeld eine ähnlich grosse Wirkung aus. Untersuchungen der Seen und Flüsse im Bereiche der heutigen "kalten" Gletscher tragen zu einem besseren Verständnis eiszeitlicher Formen und Vorgänge bei.

Lacs de barrage glaciaires et écoulement glaciaire marginal.
Investigations à l'île Axel Heiberg, archipel arctique canadien.

RÉSUMÉ

Des lacs, barrés par des glaciers arctiques de vallée (type sub polaire), et des rivières juxtaglaciaires et proglaciaires ont été étudiés dans une région montagneuse et fortement englacée au centre ouest de l'île Axel Heiberg ($79\frac{1}{2}^{\circ}\text{N}$, $90\frac{1}{2}^{\circ}\text{W}$; voir figs 1-3).

Dans cette région, les lacs de barrage glaciaires sont beaucoup plus nombreux et de plus grande dimension que dans les Alpes suisses (table 3). Quatre lacs ont été explorés plus en détail:

le Between Lake - fig. 13 et table 4,	le Ice Cave Lake - fig. 26 et table 6,
le Crusoe-Baby Lake - fig. 21 et table 5,	le Phantom Lake - figs 32, 34 et table 7.

Les lacs de cette région débordent normalement à travers la surface du glacier, s'enfonçant dans la glace qui forme le barrage, ou bien le long de la marge glaciaire, ou encore par un endroit non englacé. Après un rapide accroissement du débit jusqu'au maximum, au lieu d'une brusque décrue - comme chez des lacs de barrage à vidage sous-glaciaire des régions tempérées - le débit des lacs de barrage étudiés dans le nord canadien diminuait plus lentement (fig. 39).

Quelque 130 grammes de sédiments (poids sec) par 100 cm^2 s'accumulaient en moyenne au Between Lake pendant l'été 1963 (soit pendant un remplissage et un vidage du lac). La quantité et la granulométrie des sédiments déposés pendant une telle période peuvent varier beaucoup (figs 53-55). Des matériaux dissous et suspendus dans l'eau du Between Lake et du Between River ont aussi été analysés (tables 8, 9 et fig. 48). Les conditions particulières de la sédimentation dans des lacs glaciaires qui se vident périodiquement sont discutées.

Quelques mesurages préliminaires indiquent que même un lac assez froid peut influencer sur le régime thermique du sol et de la glace submergée (figs 65-67).

La glace non tempérée et le pergélisol empêchent l'infiltration de l'eau dans le glacier et dans le sol. L'écoulement strictement marginal prédomine, mais les eaux de fonte suivent à plusieurs endroits un cours sous-marginal (figs 72, 75, 76, 88, 89). On montre comment une rivière marginale devient sous-marginale à cause de l'écroulement de la marge du glacier (figs 23, 24, 73). À la longue l'action répétitive des eaux marginales et sous-marginales peut avoir des effets géomorphologiques considérables, même sur des versants de vallée assez raides (figs 143, 144). Quelques exemples de canyons marginaux et sous-marginaux actuels sont données.

Une brève analyse de quatre régions avec d'anciens chenaux creusés par des rivières glaciaires montre que ces traces d'érosion fluviale indiquent bien d'anciennes oscillations des glaciers et révèlent d'anciens cours d'eau marginaux, sous-marginaux et supraglaciaires; de plus, elles témoignent de l'état thermique des anciens glaciers (figs 139-142).

Des modifications remarquables dans les lits des Ermine, Between et Accident Rivers et dans la plaine alluviale de l'Expedition River eurent lieu surtout pendant les périodes de crues, causées soit par l'écoulement rapide des eaux de pluie, soit par le drainage d'un lac de barrage situé en amont (figs 93, 98-109, 112-114 et table 11).

Dans cette région de pergélisol, l'effet thermique de l'eau est très important (voir p. ex. les "niches d'érosion thermique", figs 121-123). Le mode de récession des berges de rivière composées de graviers fluvioglaciaires gelés ont fait l'objet d'une étude (figs 124-132).

À part l'indication des variations glaciaires, les lacs d'obturation glaciaire jouent le rôle d'immenses totalisateurs d'ablation glaciaire et d'enregistreurs météorologiques (fig. 138 et table 13).

Le régime des rivières en aval des lacs de barrage glaciaires est bien différent de celui des rivières glaciaires sans lacs de barrage dans leur cours. La présence de ces lacs peut concentrer l'écoulement sur peu de jours de crue avec un effet géomorphologique considérable. Comme les lacs de barrage glaciaires sont relativement nombreux à l'île Axel Heiberg, on pense que les crues causées par le vidage de ces lacs jouent un rôle important dans le façonnement des lits de rivières marginaux et proglaciaires, et, à la longue, dans le façonnement des vallées glaciaires. Une grande partie des glaciers quaternaires était non tempérée - au moins pendant certaines périodes - et de nombreux lacs côtoyaient ces glaciers. Ces lacs se vidèrent probablement d'une manière semblable à celle décrite dans ce rapport et les crues qui en résultèrent eurent un effet pareil. Une étude des lacs et rivières aux environs des glaciers arctiques actuels peut donc aider à mieux comprendre les données des glaciations quaternaires.

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- 1962: White Glacier, Thompson Glacier Region, Axel Heiberg Island, Canadian Arctic Archipelago; scale 1: 5 000
- 1962: Thompson Glacier Snout, Axel Heiberg Island, Canadian Arctic Archipelago; scale 1: 5 000
- 1964: White Glacier, Axel Heiberg Island, Canadian Arctic Archipelago (in two parts); scale 1: 5 000
- 1965: Crusoe Glacier Tongue, Axel Heiberg Island, Canadian Arctic Archipelago; scale 1: 5 000
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